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COMPARISON OF THE FABRICS OF INCLUSIONS AND THE ADJACENT INTRUSIVE ROCK

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INTRODUCTION

Fabric studies are being used more and more in interpreting the history of metamorphic areas and have been used a few times to obtain from igneous and sedimentary rocks information not available by other means. The use of such studies is by no means confined to making large-scale tectonic interpretations; they can aid in attacking many kinds of geological and mineralogical problems. For example, statistical grain orientation studies can aid in determining the origin of inclusions in igneous rocks and of included minerals¹ within individual grains of other minerals; also whether or not metamorphism has affected inclusion and host together.

Before such fabric criteria of origin can be very useful or reliable it will be necessary to make numerous studies of actual examples. This paper is such a study and has to do with three² intrusions that have numerous inclusions: (1) The Port Deposit granodiorite of Northeastern Maryland, (2) the Uncle Sam quartz monzonite porphyry, which is just southwest of Tombstone, Arizona, and (3) the Val Verde tonalite

¹ Ingerson, E., Albite trends in some rocks of the Piedmont: *Am. Jour. Sci.*, **35A**, 127-141 (1938). In this paper the following paragraph appears:

"These albite trends are most striking where mica is the included mineral, but many consist of other minerals such as quartz, epidote, or garnet. These trends of other minerals are more difficult to study quantitatively, but where they are present in the same feldspar grain with mica they appear to be parallel to the mica and presumably have the same origin and significance."

Since the appearance of this paper several quartz trends in albite grains have been measured accurately and they correspond to crystallographic planes of the albite porphyroblasts quite as closely as do the mica trends. Five of them were measured in five different albite grains and were found to be parallel to the following planes, (010), (010), (001), (110), and (110).

² The first two of these were studied by the writer. The Val Verde tonalite was studied by Dr. E. F. Osborn, who very kindly lent specimens for fabric comparison.

of Southern California. The first of these has been subjected to intense regional metamorphism since intrusion; the others have not.

PORT DEPOSIT COMPLEX

The Port Deposit granodiorite complex³ intrudes older intrusive rocks as well as schists and metamorphosed extrusive rocks. The specimens from which orientation diagrams have been prepared were collected

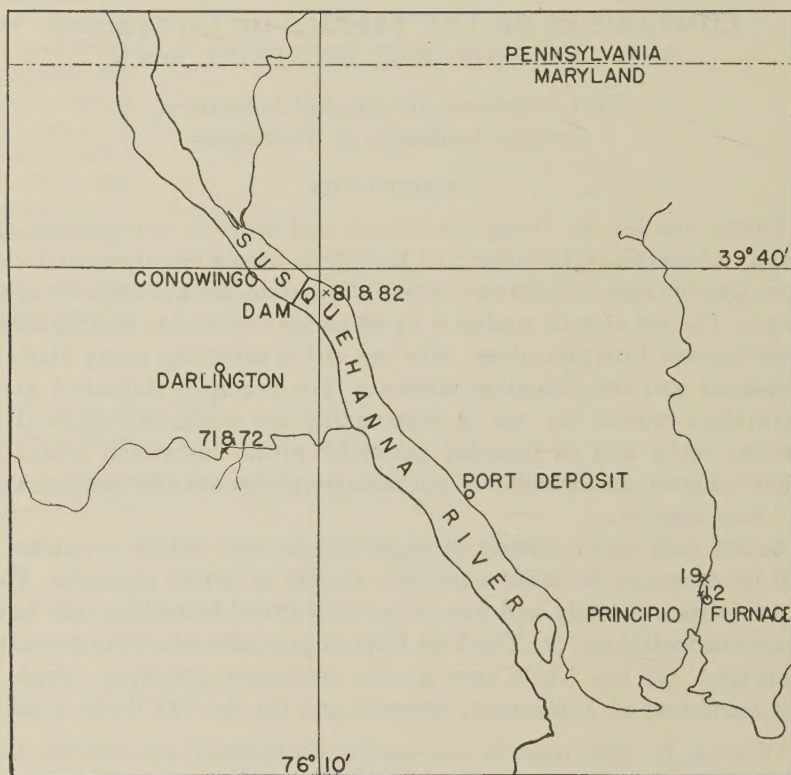


FIG. 1. Index map showing the locations of the specimens from the Port Deposit complex.

from three different localities in the area: (1) at Principio Furnace, near the contact with metadacite (12 and 19), (2) from a road cut on the south side of Deer Creek, 3 miles S 30°W from the dam (71 and 72), and (3) just below Conowingo Dam (81 and 82). See Fig. 1.

The intrusive at Principio Furnace is a light gray granitic rock. The texture is entirely metamorphic, but its intrusive nature is indicated by

³ For a description of the rocks of the Port Deposit granodiorite complex, and a bibliography on the area, see Hershey, H. G., Structure and age of the Port Deposit granodiorite complex: *Maryland Geol. Survey*, 13, 107-148 (1937).

the contacts with the country rock, and the presence of large numbers of inclusions. These inclusions are variable but in general contain essentially the same minerals as the granite. The fabrics of two inclusions from Principio Creek have been studied. No. 12 was collected just above the bridge on U. S. 40, and No. 19 from a cut bank a quarter of a mile up the creek.

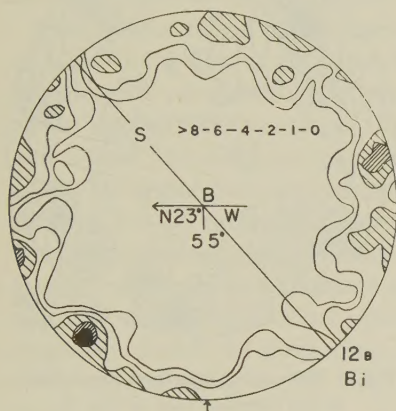


FIG. 2. Poles of 100 biotite flakes from the intrusive rock at Principio Furnace, Maryland.

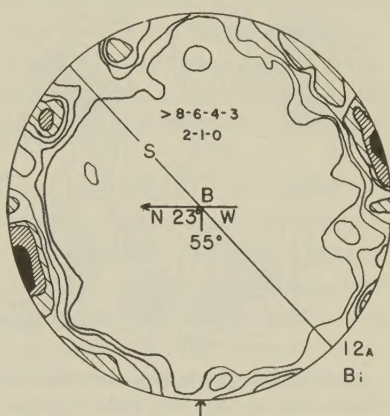


FIG. 3. Poles of 200 biotite flakes from an inclusion in the intrusive rock from which the diagram of Fig. 2 was prepared.

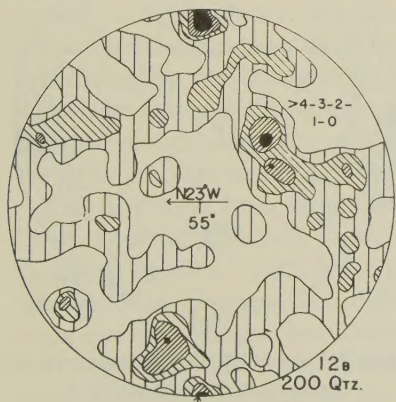


FIG. 4. 200 quartz axes from the intrusive at Principio Furnace.

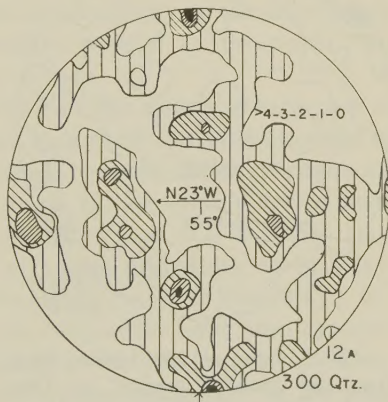


FIG. 5. 300 quartz axes from the inclusion of Fig. 3.

In No. 12 the inclusion differs from the host rock only in higher percentages of hornblende, biotite, and garnet. The foliation plane is common to host and inclusion, running across the contact with no deviation

at all. The lineation in the foliation also appears to be the same in both rocks.

Biotite and quartz orientation diagrams were prepared from a thin section cut normal to the lineation (Figs. 2-5). The biotite diagram from

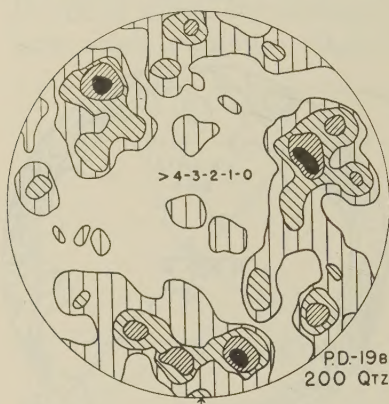


FIG. 6. 200 quartz axes from the intrusive rock up the creek from Principio Furnace, near the contact with metadacite.

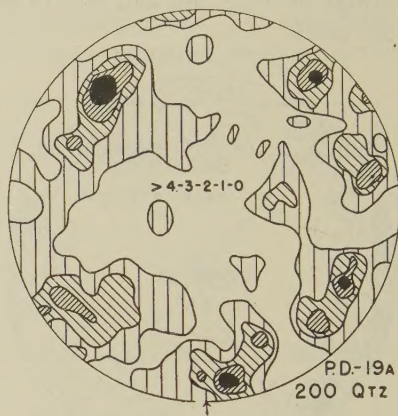


FIG. 7. 200 quartz axes from an inclusion in the intrusive rock of Fig. 6.

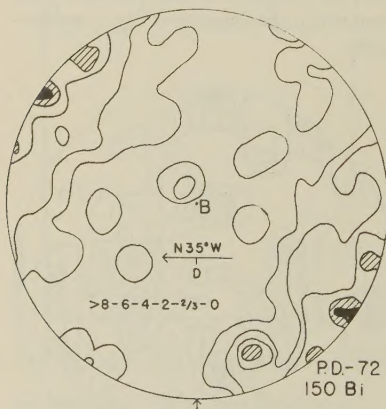


FIG. 8. Poles of 100 biotite flakes from the "pudding granite" on Deer Creek southeast of Darlington, Maryland.

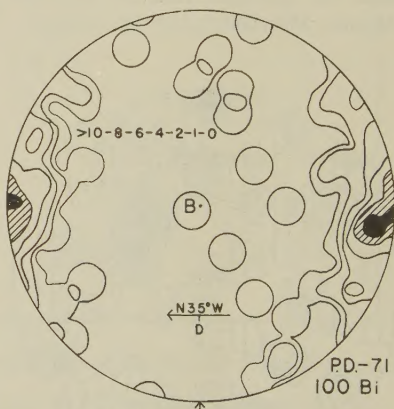


FIG. 9. Poles of 100 biotite flakes from an inclusion in the rock of Fig. 8.

the host rock, Fig. 2, shows a well developed girdle normal to the lineation. The principal maximum is normal to the *s*-plane, which is the arrangement that is to be expected. In the inclusion also the biotite

forms a girdle normal to the lineation, but the principal maximum is not normal to the megascopic *s*-plane. This indicates that host and inclusion were subjected to the same metamorphic environment after intrusion was complete. The difference in the position of the maxima is probably due to the effect of an older set of *s*-planes in the inclusion, possibly the original foliation. There is no chance for difference in orientation of thin sections, because both diagrams were prepared from the same section.

The quartz fabric of the inclusion has been even more completely re-made. Both quartz diagrams, Figs. 4 and 5, have the principal maximum on the periphery near the orientation arrow, a decided minimum extending from the center to the periphery in the NE-SW quadrants, and almost identical sub-maxima.

The intrusive rock and inclusion of specimen No. 19 are much like those of No. 12, except that they have less quartz and more hornblende. The intrusive rock contains very little biotite and the inclusion has none at all. Inclusions from this part of the complex,—near the contact with volcanic rocks,—have been called metadacite. If this inclusion was originally dacite the plagioclase has been albitized. The texture is entirely metamorphic and it may just as well have been a sedimentary rock.

This specimen does not have as definite a foliation as does No. 12, but there is a marked lineation visible in the elongate hornblende crystals, and it is parallel in host and inclusion. A thin section cut normal to the lineation (strike of thin section, N 23°W; dip 55°SW; arrow points up dip, N 67°E) shows both cleavages in a majority of the hornblende grains, indicating that they are elongate parallel to the *c*-axis. No fabric diagram was prepared from hornblende.

An orientation diagram was prepared from the quartz in each part of the thin section, Figs. 6 and 7. In each case there is a girdle normal to the lineation, indicating that the relation is the same as for specimen 12.

The specimens from Deer Creek (71 and 72) are quite different from those taken on Principio Creek. The intrusive rock (72) is a dark gray "pudding granite," with white blebs of quartz and albite up to half a centimeter in diameter. It is difficult to classify most of these rocks, because not only texture, but also mineral composition have been changed since intrusion. The most obvious change is the development of albite porphyroblasts, but much of the quartz is either secondary, or has been recrystallized, and some of the mafites may be secondary. On the basis of the present mineral composition this intrusive rock is (after Johannsen) a sodaclase-granodiorite. If the albite has replaced more basic plagioclase the rock was originally a normal granodiorite; if it has replaced potassium feldspar the rock may have been a hornblende granite.

The inclusion (71) is a fine-grained, dark-gray rock that looks, in the hand specimen, much like the "groundmass" of 72. In thin section the mineral composition is seen to be much the same except that there is much more hornblende and somewhat less quartz and feldspar. These inclusions have been called "gabbro." They are not gabbro now but may have been originally, if all the quartz is secondary and if the original feldspar was labradorite.

It is not necessary, however, to decide what these rocks were, or even what names should be given them now, in order to accomplish the present purpose, which is merely to compare fabrics of host rocks and inclusions.

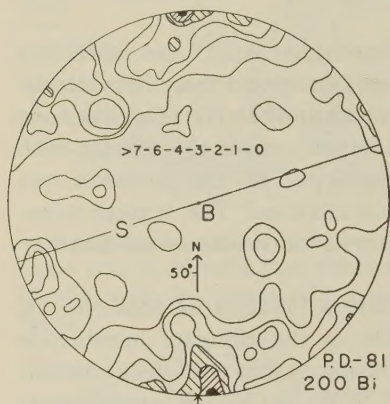


FIG. 10. Poles of 200 biotite flakes from the more acidic facies of the intrusive rock exposed below Conowingo Dam.

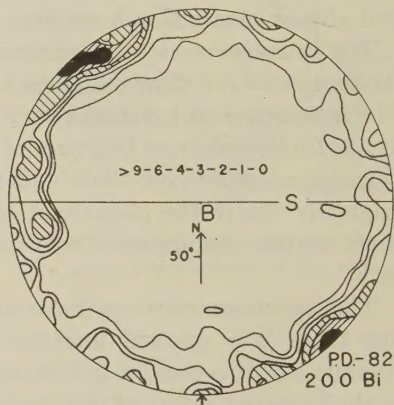


FIG. 11. Poles of 200 biotite flakes from an inclusion of quartzite in the intrusive rock of Fig. 10.

Specimens 71 and 72 appear almost structureless in the field, but there is a faint linear structure. Thin sections were cut normal to this lineation and Figs. 8 and 9 show biotite diagrams from intrusive and inclusion respectively. Each shows a partial girdle normal to the lineation, but the maxima do not coincide. This indicates, again, that the two have been affected by the same deformation after intrusion, but that response was along slightly different *s*-planes in the two rocks.

Immediately below Conowingo Dam the rocks are well exposed in the river bed at low water. Hershey⁴ has prepared a detailed map of this area showing three different intrusive rocks, an older amphibolite and a series of sedimentary schists. An inclusion of the sedimentary rock (82) and a piece of the adjacent "granodiorite" (81) were collected for the purpose of comparing the fabric of an inclusion of known sedimentary origin with that of the enclosing rock.

⁴ *Maryland Geol. Survey*, 13, Pl. XI.

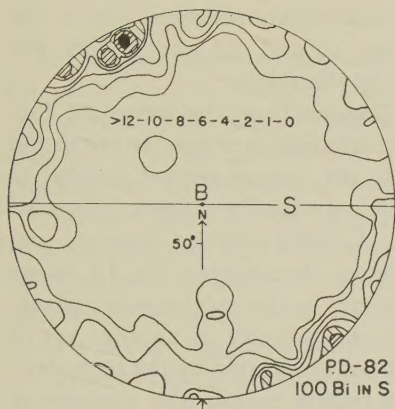


FIG. 12. Poles of 100 biotite flakes lying in the *s*-plane of the inclusion of Fig. 11. Figs. 12 and 13 are the partial diagrams that were added together to give Fig. 11.

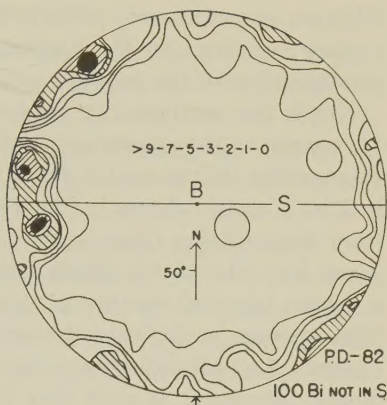


FIG. 13. Poles of 100 biotite flakes lying between the visible *s*-planes of the inclusion of Fig. 11.

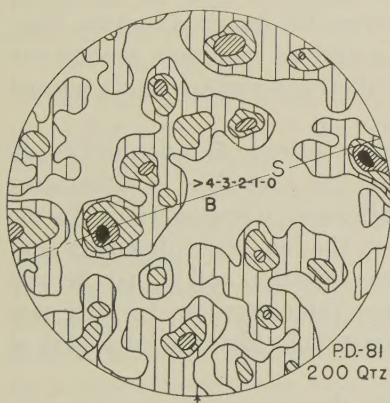


FIG. 14. 200 quartz axes from the intrusive rock below Conowingo Dam. Same thin section from which Fig. 10 was prepared.

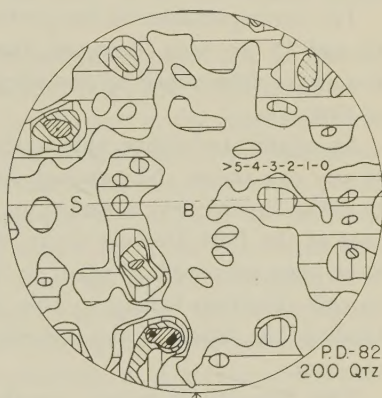


FIG. 15. 200 quartz axes from the inclusion from which the diagrams of Figs. 11 to 13 were prepared.

The inclusion is a quartzite with well-developed parting along closely spaced planes (2 to 3 mm.) containing mica. These planes may be the original bedding, but there is no independent evidence that they are. The intrusive rock is light creamy-gray and moderately fine-grained. It is high in quartz and has biotite as the dark mineral. It may well have been a granite before albitization took place. There is a rough megascopic

foliation approximately parallel to one side of the inclusion. This is the *s*-plane indicated on the diagrams (Figs. 10 and 14) and it does not correspond with the position indicated by the micas.

Nor is the maximum in the biotite diagram from the inclusion, Fig. 11, normal to the megascopic *s*-plane. In this case it is easy to see in the thin section that the mica flakes lying in the *s*-plane are not parallel to it. This is clearly shown in Fig. 12, which is a selective diagram prepared from biotite flakes lying in the *s*-plane. Another selective diagram prepared from the biotite flakes lying between the *s*-planes, Fig. 13, shows a second (double) maximum almost normal to the megascopic *s*-plane. This appears to indicate the existence and activity of another set of *s*-planes. An *s*-plane, in the position of the one indicated by the secondary double maximum, is such that it could readily explain the asymmetry of the principal maximum of the non-selective diagram with respect to the megascopic *s*-plane.

This may also be the explanation for the asymmetry of the biotite diagram from the intrusive rock, Fig. 10, with respect to its foliation. At any rate, the mica fabrics in the two rocks are very similar, and they have a common *B*-axis.

The correspondence is less perfect between the quartz diagrams, Figs. 14 and 15. In each, however, there is a decided minimum around the *B*-axis (as determined by the mica) and there is a girdle roughly normal to this axis.

As is indicated above, this study is not intended to be a detailed study of any one region. Not enough field work has been done in either area to make feasible any attempt at comprehensive interpretations. In the case of the Port Deposit rocks, however, the few fabric diagrams that have been prepared all point in the same direction and it may be well to call attention briefly to this tentative interpretation. In each case studied, the fabric of the inclusion is essentially identical with that of the host.

This identity of fabric, together with the completely metamorphic texture of all the rocks, suggests that the primary structures in all the rocks have largely been obliterated and the foliations and lineations are now largely secondary metamorphic features. Where the original *s*-planes (foliation in igneous rocks and bedding in sedimentaries) were in the proper zones they may have been preserved and even emphasized by later motions. It is difficult to see, however, how the lineations common to host and inclusions can be other than secondary, superposed structures.

UNCLE SAM PORPHYRY

It has been suggested that it is possible for an intrusive rock to impart its fabric to an inclusion by continued movement after incorporation of the inclusion, but before final solidification. In order to give full consideration to this possibility it was necessary to study an igneous intrusion, with inclusions, that had not been deformed after solidification.

Professor James Gilluly suggested that the Uncle Sam porphyry would be a compact unit for such a study. It intrudes rocks ranging in age from pre-Cambrian to Mesozoic, but in the Tombstone district it is in contact principally with quartzite and limestone of the Bisbee group. Figure 16 is a sketch map modified from Butler, Wilson, and Rasor⁵ showing the general relationships. Tombstone lies just off the northeast corner of the map, half a mile to the east.

A preliminary field study indicated to Gilluly and Cannon that the porphyry might have been intruded along a thrust fault dipping some 30° to the west. They suggested that a statistical study of the orientations of platy inclusions and of their relations to the contacts might provide a definite answer. The actual contact is exposed in very few places and in these it dips 60–75° to the west, as do the nearby shear zones in the porphyry.

The orientations of all the inclusions measured could not be plotted on the map without using an impracticably large scale, but a spherical projection of the poles of the platy inclusions and of the contacts shows the relations clearly. In Fig. 17 the dots represent poles of the inclusions, and the small circles are poles of the contact surface at the various points where the orientation could be determined. (Some of these are from the area outside of the area shown on the map.) The circles are numbered and the numbered dots (with vertical lines) are poles of inclusions collected at, or near, corresponding contact exposures. The unnumbered dots with vertical lines represent poles of inclusions collected near contacts where the attitude could not be determined. A glance at the projection shows that there is no significant preferred orientation of the platy inclusions. Even near the contacts there is slight tendency for inclusions to lie parallel to the contact surfaces, there being only 5 out of 21 inclusions measured near contacts that are within 20° of parallel to the contact surfaces. Two of the inclusions stand almost normal to the contacts, and two others make angles of over 60°, the average being 40°.

⁵ Butler, B. S., Wilson, E. D., and Rasor, C. A., Geology and ore deposits of the Tombstone District, Arizona: *Univ. of Ariz. Bull.*, 9, No. 1, Plate III (1938).

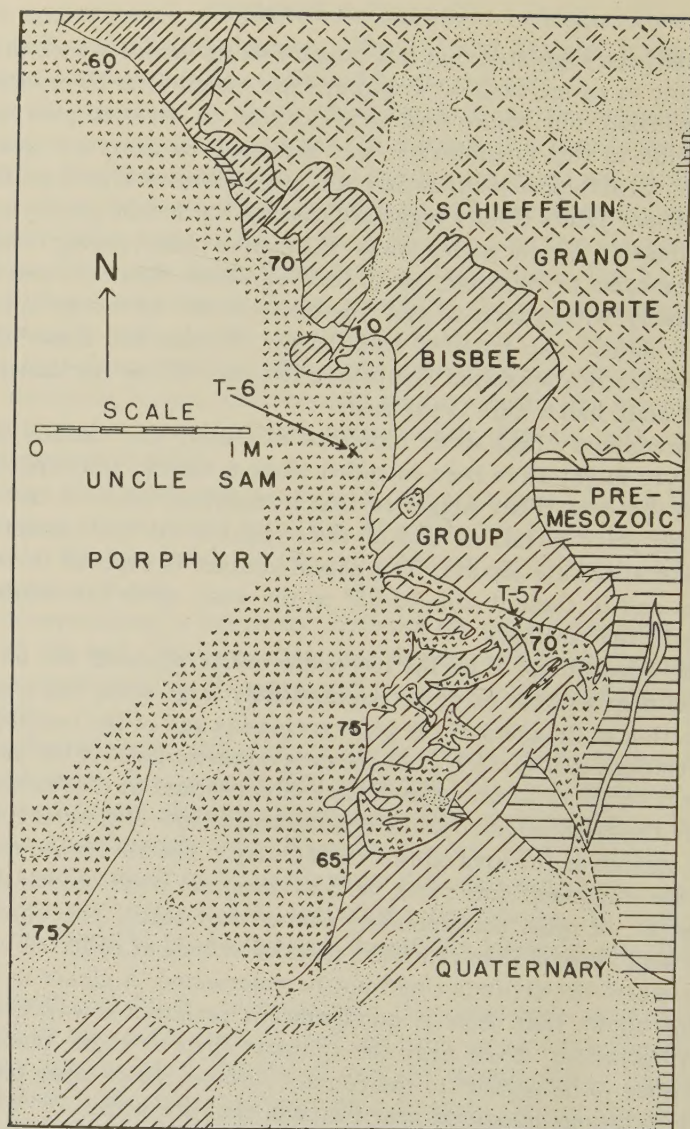


FIG. 16. Sketch map showing the eastern contact of the Uncle Sam porphyry in the Tombstone, Arizona, area. Modified after Butler, Wilson, and Rasor.

At each outcrop where inclusions were collected or measured the joint systems were measured. Prominent joint systems were also measured in other outcrops where inclusions were absent. It was thought that the attitude of the joints might possibly indicate something concerning the floor of the intrusion. The projection of the poles of the joints (Fig. 18), however, shows a more or less random distribution of the joints from vertical down to a dip of about 45° . Gilluly⁶ points out that the flatter

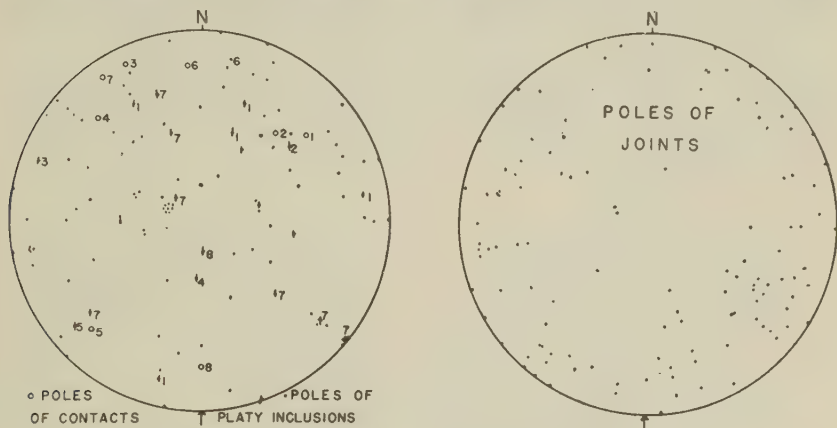


FIG. 17 (left). Equal area projection on a horizontal plane of the poles of the platy inclusions from the Uncle Sam porphyry. The poles of the contact surfaces that were measured are also shown. The numbered poles are of inclusions that lay near contacts having the same numbers. The poles through which lines have been drawn, but which are not numbered, are of inclusions that lay near contacts whose attitude could not be measured.

FIG. 18 (right). Equal area projection on a horizontal plane of the joints measured in the Uncle Sam porphyry.

joints are likely to form weathered surfaces and hence not be measured as joints. In the absence of quarries, or measurements in actual mine shafts, the steep joints are over-emphasized. Therefore, Fig. 18 does not give a true statistical picture for all the joints, but it does show the absence of well-defined systems of joints related to a shallow floor.

The detailed study of the inclusions and of the jointing does not confirm the presence of a thrust fault along which the porphyry was intruded.

The porphyry is a medium gray rock with abundant small phenocrysts (up to about 3 mm. in diameter) in a very fine groundmass. Neither foliation nor lineation is apparent. For this reason all the thin sections were cut in a horizontal plane and the orientation arrows on the diagrams point north.

⁶ Oral discussion.

The inclusions in the Uncle Sam porphyry are principally shale, limestone, and quartzite. The quartzite inclusions are the only ones in which a direct comparison of mineral orientation with that in the porphyry is possible. Quartz diagrams were prepared from three of these quartzite inclusions and from the adjacent host rock.

There is no point to discussing the individual diagrams, because they bear no apparent relation either to structures in the intrusive rock or to the flat surfaces (=bedding?) of the inclusions. Figures 19 and 20 are quartz diagrams from an inclusion and the adjacent porphyry respectively. The diagrams are not at all similar; the principal maximum of the diagram from the porphyry coincides with the largest minimum in the other diagram.

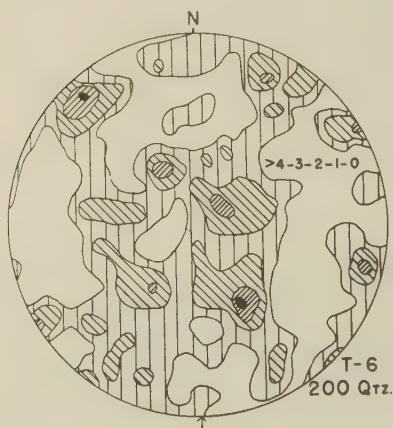


FIG. 19. 200 quartz axes from a quartzite inclusion in the Uncle Sam porphyry.

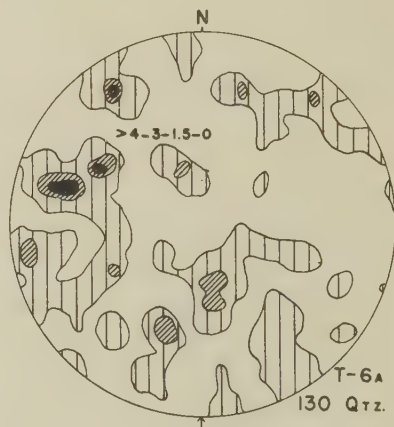


FIG. 20. 200 quartz axes from the porphyry adjacent to the inclusion of Fig. 19.

Figures 21 and 22 are from an inclusion and porphyry taken at another outcrop and show no better correspondence. The two inclusions from which the diagrams of Figs. 19 and 21 were prepared are comparatively large ($3 \times 4 \times 6$ cm. and $8 \times 15 \times 20$ cm.); in one of the thin sections of porphyry (T-57a) there is a cross-section of a small quartzite inclusion, its greatest diameter being less than 4 mm. There has been enough reaction so that the boundary between the inclusion and porphyry is not sharp, being apparent only by the predominance and equigranularity of quartz in the inclusion, yet the quartz fabric (Fig. 23) has not been affected and shows no more correspondence to that of the porphyry than does that of the large inclusion.

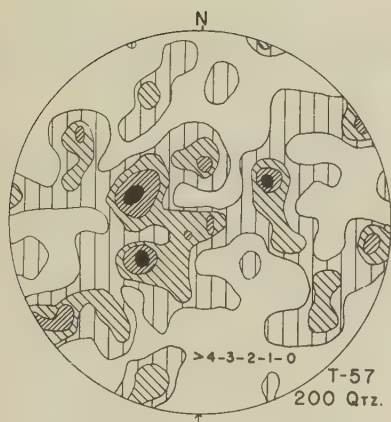


FIG. 21. 200 quartz axes from another quartzite inclusion in the Uncle Sam porphyry.

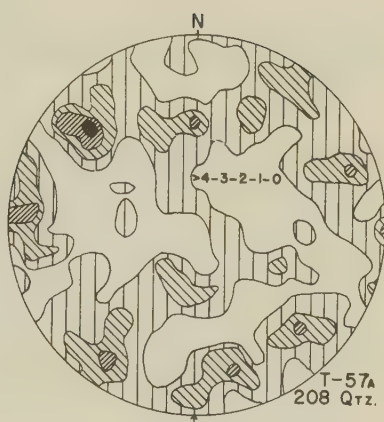


FIG. 22. 200 quartz axes from the porphyry adjacent to the inclusion of Fig. 21.

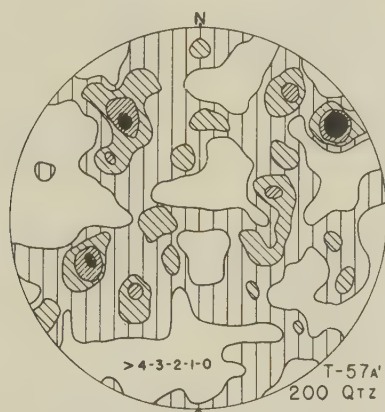


FIG. 23. 200 quartz axes from a small quartzite inclusion in the thin section from which the diagram of Fig. 22 was prepared.

For this intrusion, then, the fabric of the xenoliths is quite different from that of the intrusive rock. This must not be taken as a general rule for unmetamorphosed intrusions, however, until other examples have been studied. Inclusions incorporated in deeper-seated intrusions would have a much better chance to be heated to the temperature of the magma before its final consolidation. The higher temperature, together with

continued motion of the crystallizing magma, might change the fabric of inclusions and could conceivably make it almost identical with that of the host.

VAL VERDE TONALITE

The tonalite intrusions studied by Hurlbut⁷ and Osborn⁸ may be examples of the action of such deep-seated intrusions on included material. However, the depth of intrusion of these bodies is not known, and it is by no means certain that the inclusions are xenoliths.

Hurlbut bases his interpretation of the inclusions as xenoliths on mineralogical and chemical similarities between the inclusions and an older gabbro in contact with the tonalite, and differences between these

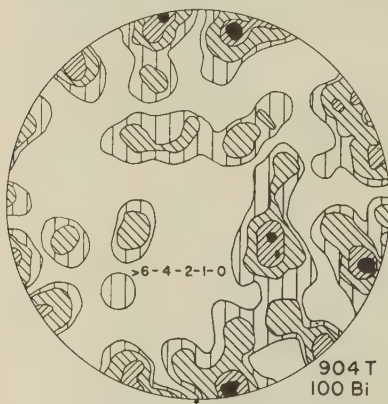


FIG. 24. Poles of 100 biotite flakes from the Val Verde tonalite. The diagrams of Figs. 24-29 were all prepared from parallel thin sections cut from a single hand specimen showing both inclusion and host.

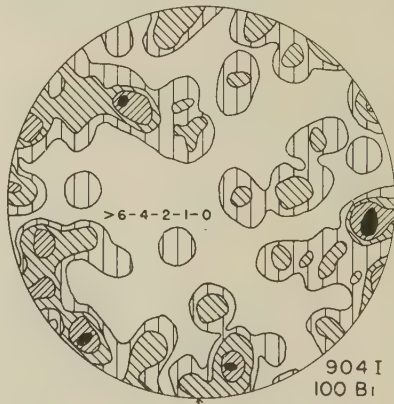


FIG. 25. Poles of 100 biotite flakes from an inclusion in the hand specimen from which the diagram of Fig. 24 was prepared.

rocks and the tonalite. However, these differences are such as would be expected if the inclusions are early segregations, i.e. "autoliths." Moreover, he points out that the mineral orientation is the same in inclusions and tonalite, but shows a *greater* preferred orientation in the inclusions than in the host. This relation might be expected in an autolith. In a xenolith the degree of preferred orientation might approach that of the host, but it is difficult to imagine its going beyond that of the host.

⁷ Hurlbut, C. S., Jr., Dark inclusions in a tonalite of Southern California: *Am. Mineral.*, 20, 609-630 (1935).

⁸ Osborn, E. F., Structural petrology of the Val Verde tonalite, Southern California: *Bull. Geol. Soc. Am.*, 50, 921-950 (1939).

Osborn found similar orientations of quartz and biotite in inclusions and host, but did not make a comparison between a given inclusion and the adjacent intrusive rock.

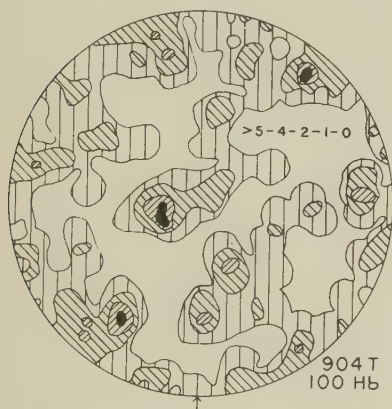


FIG. 26. 100 hornblende c-axes from the Val Verde tonalite.

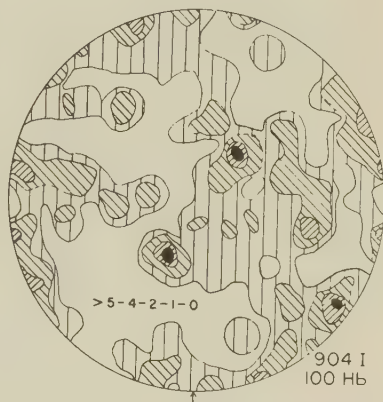


FIG. 27. 100 hornblende c-axes from an inclusion in the Val Verde tonalite.



FIG. 28. 200 quartz axes from the Val Verde tonalite.

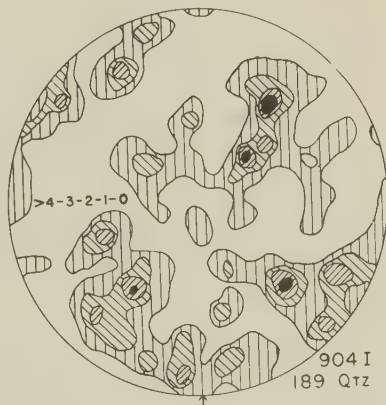


FIG. 29. 189 quartz axes from an inclusion in the Val Verde tonalite.

He was kind enough to lend some of his specimens for the present study, in which inclusion and host occur in a single hand specimen. In these specimens the foliation and lineation are but poorly developed and the same degree of preferred orientation is not to be expected as in more highly foliated specimens. Thin sections were cut normal to the rough lineation, and biotite, hornblende, and quartz diagrams were

FORMATION AND STABILITY OF MUSCOVITE IN ACID SOLUTIONS AT ELEVATED TEMPERATURES

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INTRODUCTION

The experiments described below are a few of a larger number undertaken to determine the stability of layer silicates in HCl solutions at temperatures between 300° and 400°C. The present paper is of a preliminary nature only. Enough information has been gathered, however, to be of interest in connection with similar studies by Noll¹ and Norton.² Especially Noll's conclusions are significant and have been confirmed by the writer's experiments in most points. It seems, however, that the formation and stability of such minerals as sericite and kaolinite should be greatly affected by the concentration of alkalis, not only in basic solutions, as has been known for many years, but also in acid solutions. The writer found that the concentration of KCl in his experiments with HCl solutions was of great importance in stabilizing muscovite which otherwise would decompose to kaolinite or pyrophyllite.

EXPERIMENTS

As far as the writer knows Schwarz and Trageser³ are the only ones who have used HCl solutions in bomb experiments at temperatures around 300°C. They speak of steel bombs containing platinum and silver crucibles but do not mention how they protected the steel from the corrosive action of the gases. In the present experiments gold-lined bombs, which have been described a number of times, were used. The bomb has a volume of 50 cc. In all experiments it was filled with 25 cc. of solution. The air above the liquid was displaced by CO₂ (at atmospheric pressure and room temperature), and the bomb was sealed by a sheet of gold which was pressed against the gold flange of the lining. After the experiment a circular hole was cut in the gold sheet. The cooling of the bomb took about 2 hours. The heating of the bomb was carried out in a large oven-like electric furnace controlled by a Leeds and Northrop recording potentiometer. The thermocouple was as close to the

¹ Noll, W., Über die Bildungsbedingungen von Kaolin, Montmorillonit, Sericit, Pyrophyllit und Analcim: *Mineral. Petrograph. Mitt.*, **48**, 210-247 (1936). This is a summary of several papers on the same subject by Noll.

² Norton, F. H., Hydrothermal formation of clay minerals in the laboratory: *Am. Mineral.*, **24**, 1-18 (1939).

³ Schwarz, R., and Trageser, G., Über die künstliche Umwandlung von Feldspat in Kaolin: *Zeits. anorg. Chem.*, **215**, 190-200 (1933).

bomb as possible. Fluctuations in temperatures of the furnace which amounted to a maximum of $5^{\circ}\text{C}.$ \pm were minimized by the considerable weight of the bomb. The materials used for the experiments were:

Muscovite in large books from the Hugo Mine, Black Hills, South Dakota. The mica was filed to about 200 mesh size.

Kaolinite from Brooklyn, N. Y. The same material was used in the determination of the structure of kaolinite. For its analysis see: Gruner, J. W., *Zeits. Krist.*, **83**, 75 (1932).

Aluminum hydroxide (Merck). This gives an x -ray pattern of bayerite and contains 37% H_2O .

"*Silicic acid*" (C. P. Mallinckrodt). It contains 10.36% H_2O = approx. $\text{SiO}_2 \cdot \frac{1}{2}\text{H}_2\text{O}$.

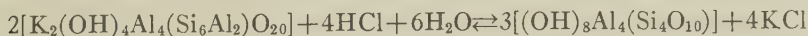
Two hundred milligrams of muscovite and kaolinite were used in each experiment. The amounts of the hydroxides used for synthesis of the minerals are stated for each experiment.

After completion of most experiments, 20 cc. of the filtered solution were titrated with NaOH to find the remaining acidity, and evaporated for SiO_2 analysis by hydrofluorization. The powders were x -rayed in precision cameras with radii of 57.3 mm. Unfiltered Fe radiation was used. Very good sharp patterns were obtained in all experiments described. To reproduce the line measurements here would be just a waste of space as none of the diagrams are ambiguous. They seem to be much better than those of Norton,⁴ which perhaps is due to more thorough and complete recrystallization in HCl solutions.

Besides the nine experiments listed in Table 1, five others are tabulated in Table 2. The pressures in the bombs were approximately 85 atm. at 300°C . for experiments 106, 111, and 126, and 300–400 atm. in experiment 154 at 400°C . The addition of KCl would cause, however, a lowering of pressures.

DISCUSSION OF RESULTS

Experiments 111 and 126 show that muscovite is not stable in HCl solutions at 300°C ., provided the concentration of KCl stays below a certain limit. This limit seems to be indicated by the results in 106. Apparently some muscovite changed to kaolinite and when KCl had reached a certain concentration with respect to the concentration of HCl the conversion stopped. The following equation may be written:



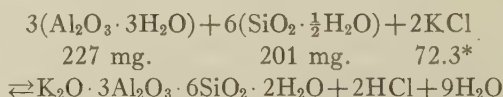
Experiments 129, 125, 141, 131, and 147 all indicate that excess of KCl prevents the change of muscovite to kaolinite. On the other hand,

⁴ *Op. cit.*

the reaction will not go from right to left in acid solutions at 300°C. as shown by experiments 130 and 138 even with a very large excess of KCl.

The synthesis of muscovite from the hydroxides does not succeed even though a very large excess of KCl is present as shown by experiment 149 which confirms Noll's⁵ contention. On the other hand, at the higher temperature of 400°C. muscovite will form from the hydroxides in acid solutions as shown by experiments 153 and 157. No pyrophyllite was obtained as Noll⁶ expects should form under these conditions.

Experiments 153 and 157 differ from each other in the amount of $\text{Al}(\text{OH})_3$ used. It was thought at first that an amount of $\text{Al}(\text{OH})_3$ sufficient to "neutralize" the HCl in the solution, as well as that formed by the decomposition of KCl, would be necessary to cause the reaction to go in the desired direction. This extra amount caused the formation of boehmite and evidently was not necessary as indicated by experiment 157, in which the weighed-out amounts of the hydroxides correspond to the following equation, the amount of the expected muscovite being about half a gram:



It may be of interest to mention that in an experiment such as 149 in which KF was substituted for KCl, leucite was the result. The original acidity in this case was N/0.103, but the solution reacted basic at the end of the experiment. That the reaction can go from right to left in the first equation at 400°C. is definitely established by experiment 155.

It would be of considerable interest to know the temperature-concentration diagram of this reaction more precisely. Since pyrophyllite is not common, it may be assumed that under most natural conditions existing at or above 400°C. the K ion is present in appreciable concentrations in an hydrothermal environment causing sericitization. This probably applies especially to neutral and acid solutions. In basic ones minerals with higher alkali content like the potash feldspars may even predominate. As the temperature decreases, the stability of the potash bearing minerals will depend primarily upon the K ion concentration in the solutions coming in contact with the minerals, and secondarily upon the acidity, or hydrogen ion concentration.

⁵ *Op. cit.*, 226.

⁶ *Op. cit.*, 227.

* Large excess used.

TABLE 1. HYDROTHERMAL EXPERIMENTS AT 300°C. IN HCl SOLUTIONS
Amount of mineral in each experiment=200 mg.

No.	Original mineral	Grams of KCl	Original acidity	Resulting acidity	Time in days	SiO ₂ in solution	Conversion product
106	muscovite	none	N/0.103	n.d.	10	n.d.	muscovite and some kaolinite
111	muscovite	none	0.350	n.d.	8	n.d.	kaolinite
126	muscovite	none	0.175	0.138	10	920 pt. p.m.	kaolinite
129	muscovite	3.0 g.	0.350	0.264	8	1075 pt. p.m.	muscovite
125	muscovite	1.0 g.	0.350	n.d.	8	n.d.	muscovite
141	muscovite	0.5 g.	0.175	0.138	10	805 pt. p.m.	muscovite
131	muscovite	0.5 g.	0.350	0.269	8	815 pt. p.m.	muscovite
147	muscovite	0.25 g.	0.175	n.d.	14	n.d.	muscovite
130	kaolinite	5.0 g.	0.103	0.085	10	940 pt. p.m.	kaolinite
138	kaolinite	5.0 g.	0.050	0.042	15	950 pt. p.m.	kaolinite and some boehmite

TABLE 2. HYDROTHERMAL EXPERIMENTS AT 300°C. AND 400°C. IN HCl SOLUTIONS

No.	Material placed in bomb	Original acidity	Time in days	Conversion product	Temp.	Remarks
154	200 mg. muscovite	N/0.103	5	muscovite	400°	About half of sol. escaped. Acidity 0.073
153	317 mg. Al(OH) ₃ 201 mg. SiO ₂ · xH ₂ O 2000 mg. KCl	0.050	5	muscovite and boehmite	400°	Acidity 0.058 SiO ₂ =570 pt. p.m.
149	As in 153	0.050	13	kaolinite and boehmite	300°	Acidity 0.039 SiO ₂ =822 pt. p.m.
157	227 mg. Al(OH) ₃ 201 mg. SiO ₂ · xH ₂ O 2000 mg. KCl	0.050	5	muscovite	400°	Sol. lost at end of exp.
155	200 mg. kaolinite 2000 mg. KCl	0.103	5	muscovite	400°	About $\frac{1}{3}$ of sol. recovered.

CONCLUSIONS

Experiments show that the formation of muscovite or sericite does not occur in HCl solutions even in the presence of a large excess of KCl, at temperatures around 300°C. At 400°C. muscovite will form, however, in HCl solutions from aluminum hydroxide, silica and KCl. Also kaolinite will be converted to muscovite at this temperature in the presence of an excess of KCl. The presence of KCl will stabilize muscovite in HCl solutions at temperatures around 300°C. under conditions which do not permit its formation but would ordinarily cause the formation of kaolinite. One may conclude, therefore, that in hydrothermal solutions the concentration of K ions is of primary importance in connection with their reactions upon potash-bearing minerals. Lack of K ions will cause their decomposition; above probably 350° to 375°C. pyrophyllite may form, below this temperature, kaolinite. The upper and lower temperature limits of these minerals are unknown at present.

MARSHITE AND OTHER MINERALS FROM CHUQUICAMATA, CHILE

O. W. JARRELL, *Harvard University, Cambridge, Mass.*

A detailed account of the mineralogy and paragenesis of the copper deposit at Chuquicamata, in the arid Atacama Desert of north Chile, has recently been published by M. C. Bandy (1938). While Dr. Bandy was carrying out his investigation at Harvard University, the present writer was employed as a staff geologist at the mine. During that time and since the publication of Bandy's paper, several additional minerals have been found at Chuquicamata. Of these, salesite, a new mineral, has recently been described (Palache and Jarrell, 1939), and the present paper is intended to list the others, and thus make as complete as possible the description of the mineralogy of this remarkable deposit.

MARSHITE

CuI

Marshite, the natural cuprous iodide, was discovered by C. W. Marsh (1893) in the oxidized zone of the lead, zinc, and silver deposit at Broken Hill in the arid region of western New South Wales. Spencer and Prior (1901) published a complete description of the crystals from Broken Hill. Aminoff (1922) has investigated the crystal structure, and various other workers have prepared or studied synthetic material.

In 1937, some crystals intimately associated with atacamite were found at the south end of the oxidized ore body at Chuquicamata. Because of their color, isotropic nature, and high index of refraction, they were suspected of being marshite. Dr. Berman later confirmed this identification. Recently Mr. Lester Zeihen of the mine staff at Chuquicamata has sent additional specimens to Harvard University.

Crystallography. Marshite is isometric hextetrahedral. Professor Palache examined the crystals and found that the principal faces present are the cube (100) and tetrahedron (111), which, due to oscillation, give the crystals a striated appearance, as illustrated in Fig. 1. The zone between the cube and tetrahedron is rounded due to the effect of tristetrahedral faces. Several crystals were measured, one of which yielded definite signals from the two forms $m(113)$ and $\beta(223)$, shown in Fig. 2. The striated zones gave weak reflections from other forms, none of which however is assured. Spencer (1901) has commented upon the similarity of the crystallography to that of sphalerite, and Aminoff (1922) has shown the crystal structure of the two minerals to be comparable.

Physical properties. The crystals, which range up to several millimeters across, are transparent and have an adamantine luster. When first found, they are colorless or light honey-colored. Some of them were kept for about a year at Chuquicamata, and showed no sign of change. Recently, however, some of the crystals that had been kept in darkness at Cambridge, but not protected from the air, have become salmon to reddish colored, although otherwise remaining intact. At the same time other crystals, kept in glass-stoppered bottles, but exposed to light, have remained colorless; hence the change to a reddish color seems to be promoted by contact with humid air, rather than by exposure to light.*

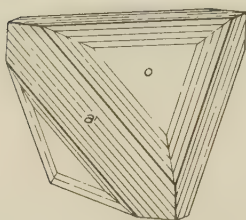


FIG. 1. Marshite.

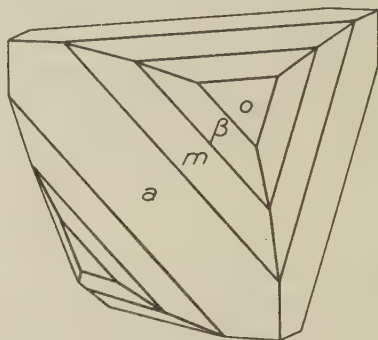


FIG. 2. Marshite.

The streak of the Chuquicamata material is only faint yellow as contrasted with the bright yellow streak that Spencer (1901) considered so striking.

Mr. Zeihen first reported that marshite was fluorescent. The color under the ultraviolet light is dark red.

The average of several determinations of the density on the microbalance gave a value of 5.68. The density calculated from Aminoff's dimensions for the unit cell is 5.60.

A small prism was cut and polished on one of the crystals, and the indices of refraction, shown in Fig. 3, for various wave lengths were determined on the one-circle goniometer by the minimum deviation method with a monochromatic illuminator. The indices are slightly lower than those obtained by Spencer on the Broken Hill material, but this small discrepancy is perhaps explained by the slight difference in chemical composition of the marshite from the two localities. The exceedingly high dispersion, exceeding that of diamond, is noteworthy.

* It has been suggested that the reddish discoloration may be due to a chemical reaction with the enclosing paper. This has not, as yet, been verified.

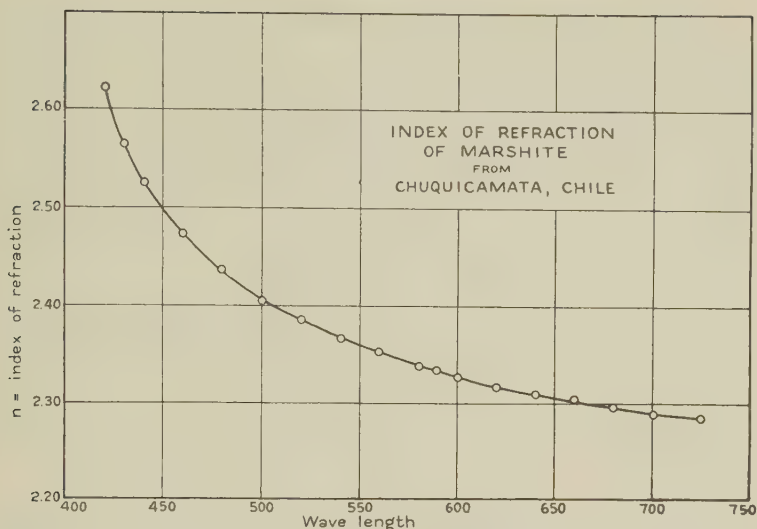


FIG. 3

Chemical composition. The analyses of the Broken Hill and Chuquicamata material, together with the ideal composition of CuI, are given below.

COMPOSITION OF MARSHITE

	1	2	3
Cu	33.01%	32.35	33.37
I	66.67	65.85	66.63
Ag	None	1.19	
Cl	0.33	—	
	100.01	99.39	100.00

1. Marshite from Chuquicamata, Chile. F. A. Gonyer, analyst.
2. Marshite from Broken Hill, N.S.W. G. T. Prior, analyst (1902).
3. Ideal composition of CuI.

For the analysis of the marshite from Chuquicamata, it was necessary to eliminate numerous tiny crystals of atacamite. This was done with the electromagnet, and before the analysis was made the sample was estimated to contain less than 1% atacamite. The amount of chlorine present in the analysis would demand about 5% of atacamite. In addition, the presence of such a large quantity of atacamite would require more copper and less iodine than the analysis shows, while in its present form the Cu-I ratio of the analysis agrees well with the ideal composition of CuI. It therefore seems best to consider most of the chlorine of the

Chuquicamata material to be present in the mineral itself, and not due to mechanically admixed atacamite.

Upon heating in the closed tube, purple fumes of iodine are given off, which condense higher up the tube as purple crystals of iodine.

Occurrence. All the specimens come from the south end of the open pit, within 12 to 24 meters of the original surface. The marshite, with atacamite, lines fractures in the sericitized and kaolinized granodiorite. In this section of the mine, antlerite, otherwise the principal oxidized copper mineral at Chuquicamata, is not stable.

The marshite crystals penetrate the atacamite and perch upon it. Probably the deposition of the minerals was essentially contemporaneous, although some of the atacamite formed after the marshite. The atacamite does not contain any appreciable amount of iodine, for its indices of refraction are normal, and no iodine is given off during heating in a closed tube.

It is interesting to note that on several occasions after blasting operations, small clouds of purplish smoke, which was shown to contain iodine, have been observed at Chuquicamata. Examination of the rock from which this gas came revealed no visible marshite or salesite. Dieseldorf (1899) has shown that several copper deposits in the arid portion of New South Wales similarly contain small quantities of iodine, although it is not present in sufficient quantities for its mineralogical nature to be recognized.

OLIVENITE



At rare intervals specimens of olivenite have been found at Chuquicamata. Excellent crystals were measured by Professor Palache and yielded the forms $m(110)$, $v(101)$ and $e(011)$, shown in Fig. 4. The following optical properties were measured on specimens sent by Mr. Zeihen:

$\alpha = 1.780$	Biaxial positive
$\beta = 1.820$	2V near 90°
$\gamma = 1.865$	$r < v$, strong

Olivinite, when found here, has always been as small crystals embedded in porous aggregates of a green arsenate, which approximates chenevixite in composition. Occasionally small amounts of an unidentified blue arsenate (?) are also present.

In previous descriptions of Chuquicamata, arsenates are stated to be relatively rare in the oxidized ore. Actually chenevixite is a common mineral in the center of the pit, and is occasionally found elsewhere. Specimens of it and the other arsenates that still contain remnants of enargite show that, where primary enargite is exposed to oxidation, copper arsenates will form in situ.

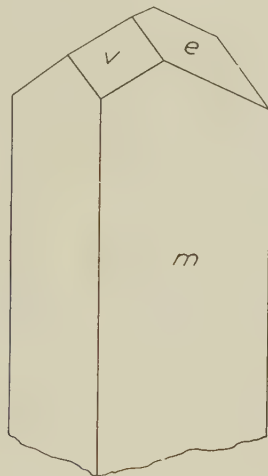


FIG. 4. Olivenite.

LIBETHENITE
 $\text{Cu}_3\text{P}_2\text{O}_8 \cdot \text{Cu}(\text{OH})_2$

A few specimens of dark olive-green crystals of libethenite perched on white, extremely kaolinized granodiorite were found on the west side of Bench E-4. This immediate area of kaolinized granodiorite contains sufficient copper to be classified as ore, although most of the rock in the vicinity is iron stained and barren of copper. Professor Palache first recognized the identity of the crystals by crystallographic measurements. The crystals show a combination of unit prism $m(110)$ and unit pyramid $s(111)$ about in the proportions of Fig. 5. The measurements are very close to those of crystals from the older known localities:

	No. of faces	Mean measured angles		Calculated	
		ϕ	ρ	Φ	ρ
{110}	3	46°13'	90°00'	46°10'	90°00'
{111}	4	46 16	45 24	46 10	45 23

These crystals are apparently the first to be described on which there is no trace of the dome $e(011)$, usually more prominent than the pyramid.

The optical properties of the crystals, which agree well with previously described material are:

$$\alpha = 1.702$$

$$\beta = 1.743$$

$$\gamma = 1.785$$

Biaxial negative

$$2V = \pm 80^\circ$$

$r > v$, strong

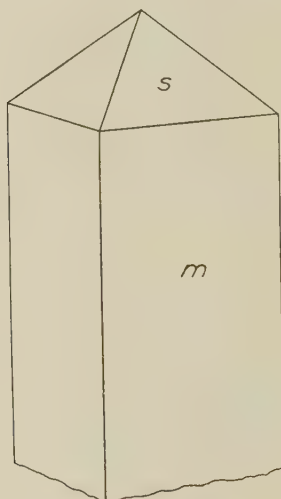


FIG. 5. Libethenite.

With the exception of turquoise, libethenite is the only phosphate so far found in the oxidized ore. Its rare occurrence can only be regarded as exceptional, and it cannot be fitted into any general paragenetic sequence for the deposit.

DARAPSKITE



Wetzel (1928) briefly noted that darapskite occurred at Chuquicamata. Actually it is a common mineral in certain zones within a few meters of the surface.

The following distinctive optical properties make it easily recognizable in aggregates with other minerals:

$$\alpha = 1.390$$

$$\beta = 1.481$$

$$\gamma = 1.488$$

Biaxial negative

$$2V = \pm 25^\circ$$

$r > v$, strong

The material available proved to be too intimately admixed with other minerals to permit of separating a pure sample for analysis.

Darapskite is found chiefly in veins up to six inches wide, intimately mixed with kroehnkite and bloedite. Good crystals have not been found. In some cases mirabilite or epsomite are also present in these veins. Both of the latter minerals are crystalline, colorless, and glassy when first broken out from the rock, but even in that dry air they break down to a white powder within a few minutes. Shellac or varnish are of little use in halting this process.

The presence, on the hillside approximately 2000 feet above the floor of the Calama Basin, of nitrate and iodine, both characteristic of the nitrate deposits in the basins of the Atacama desert, is of interest in connection with the origin of these deposits. After the sulphate radical, nitrate is the most abundant acid constituent of the groundwater at Chuquicamata.

WULFENITE



Mr. Zeihen recently recognized wulfenite in specimens from the north end of Bench E-4. It is found in dark gray masses, distinctive because of their adamantine luster, and associated with fine-grained antlerite. The optical properties are:

$$\begin{array}{ll} \text{uniaxial negative} & \epsilon = 2.30 \pm 0.02 \\ & \omega = 2.38 \pm 0.02 \end{array}$$

Wet tests for molybdenum were also obtained.

These specimens came from what has been recognized as an area of primary molybdenite mineralization, in which most of the molybdenum was retained as lindgrenite during oxidation. Minor amounts of galena have also been recognized in polished sections of the ore.

ACKNOWLEDGMENTS

Professor Charles Palache and Dr. Harry Berman have aided the writer in the preparation of this paper. Mr. George Switzer kindly helped in the preparation of the illustrations.

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SYMMETRY OF PHOSPHOSIDERITE

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Austin, Texas*

INTRODUCTION

Prior to the description of crystals from Sardinia by M. de Angelis (1) in 1926, phosphosiderite was supposed to exhibit the external symmetry of a substance belonging to the orthorhombic holohedral class. This mineral had been investigated by H. Laubmann and H. Steinmetz (2) as well as the authors who originally described the mineral, W. Bruhns and K. Busz (3). More recently phosphosiderite has been described from Kirunavara, Sweden, by R. Koechlin (4).

De Angelis demonstrated that the symmetry is monoclinic on the basis of three different types of evidence:

(1) The etch figures on the basal pinakoid do not exhibit planar symmetry with respect to a plane normal to the a axis, nor axial symmetry with respect to the c axis (Fig. 4).

(2) The goniometric measurements cannot be reconciled with orthorhombic axes of reference, but are quite amenable to monoclinic axes with $\beta = 89^\circ 24'$. Certain prominent faces are not repeated as required by orthorhombic symmetry.

(3) The optical data are not consistent with orthorhombic symmetry, because $X \wedge c > +3^\circ$.

Before the appearance of the work by De Angelis, W. T. Schaller (5) had pointed out the marked similarity between the compositions, crystal habits and axial ratios of phosphosiderite and metavariscite [formerly called variscite (6)] and he had suggested the possibility of an isodimorphous series:

Octahedral Habit	Tabular Habit
Scorodite	—
Strengite	Phosphosiderite
Variscite	Metavariscite

P. Kokkoros (7) has recently shown that scorodite and strengite are isostructural and the writer has found that variscite produces a powder diffraction pattern sufficiently similar to indicate that it also belongs to this group.

It seemed desirable to investigate phosphosiderite further by means of x -ray methods in order to ascertain whether or not the symmetry is monoclinic¹ and, if possible, to determine its structure.

¹ Although the work of De Angelis gives the appearance of care and precision, a question might arise regarding the identity of the material from Sardinia with that from Pleystein and Eiserfeld. The habit of the mineral from Sardinia is somewhat different and a different optical orientation is reported.

The crystals examined in this study are from Pleystein, Bavaria. They exhibit the forms described by Laubmann and Steinmetz (2), but a number of them show a curved, striated face which is apparently not repeated as would be required by orthorhombic symmetry. Frequently this face, or combination of faces, is quite prominent. Portions of some of the crystals and a few entire crystals are clear and pale rose colored, but many are quite cloudy and bluish-green. The coloration of these crystals is caused by the presence of many inclusions. Because of the prevalence of rounded and striated faces, the crystals do not lend themselves to accurate goniometric measurement. A considerable number of them are twins.

ACKNOWLEDGMENTS

The writer is indebted to the faculty in physics for the use of their laboratories and apparatus. He is particularly indebted to Professor M. Y. Colby for reading the manuscript and for numerous helpful suggestions. Professors F. L. Whitney and Arnold Romberg kindly assisted with the translation of the work by De Angelis. Dr. W. F. Foshag, of the U. S. National Museum, kindly furnished the specimens of phosphosiderite and strengite, scorodite and variscite. Mr. Lynn Gardiner, of the University of Minnesota, kindly prepared the powder diffraction patterns of these specimens.

EXPERIMENTAL DATA

Comparatively little difficulty was encountered in obtaining a Laue photograph in the direction of the b axis. A photograph parallel to a was obtained with considerably greater difficulty because of the thin tabular habit of the crystals and the presence of innumerable inclusions. Both of these photographs indicate the centrosymmetric group C_2^h and demonstrate the proximity of β to 90° (Figs. 1 and 2).

A series of oscillation photographs was prepared using MoK radiation. Intensity estimates, which were obtained from these photographs, greatly facilitated the interpretation of the powder diagram (Table 1). Indeed, many of the ambiguities which would otherwise arise in the interpretation of the powder diagram can only be eliminated through consideration of the data obtained from the Laue and oscillation photographs.

Specimens from Sardinia and Eiserfeld were not available for the present study, but a powder diffraction pattern of the specimen from Pleystein is given and a comparison with samples from other localities can be obtained readily on that basis.

However, the material which De Angelis examined must have been phosphosiderite, in spite of the difference in habit, and numerous reasons for this conclusion will appear in the description of the present results.

The dimensions of the unit cell were ascertained by comparison with a pattern of calcite. The values obtained are:

Absolute	Ratios	Ratios (De A.)
$a_0 = 5.30 \text{ \AA}$	0.541	0.5449
$b_0 = 9.79$	1.	1.
$c_0 = 8.67$	0.886	0.8968
(all $\pm 0.01 \text{ \AA}$)	$[\beta = 89^\circ 24']$	$\beta = 89^\circ 24'$

The methods employed are not sufficiently accurate to permit a precise determination of the angle β when the deviation from 90° is of this order of magnitude. The result obtained by goniometric measurement (1) was used in the calculation of the interplanar distances (Table 1) and the agreement is most excellent.

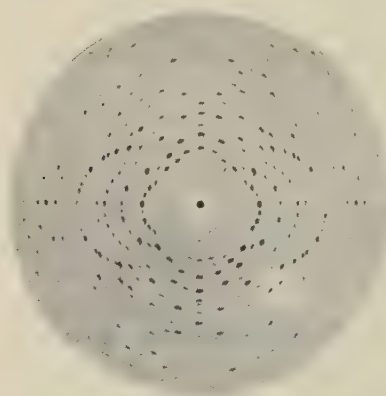


FIG. 1. Laue photograph with incident ray nearly parallel to the b axis, showing axial symmetry of spots, but not planar symmetry.

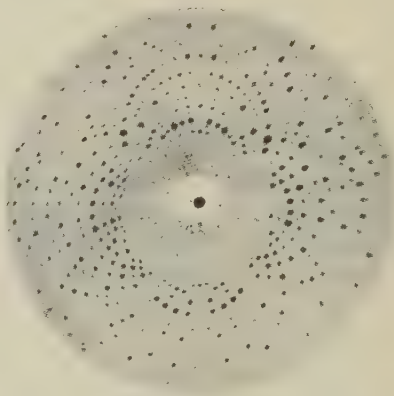


FIG. 2. Laue photograph with incident ray slightly inclined to the a axis.

It was not possible to obtain a sample sufficiently pure to completely exclude extraneous lines from the powder diffraction diagram, and several of the lines given in Table 1 were contributed by strengite, as indicated by *Str.*—the subscripts indicating the relative intensities of these lines in the strengite pattern. One other impurity was also present and three extraneous lines are probably due to this substance, the lines of which are marked *Imp.* This substance was not identified but some of the more likely possibilities were eliminated, including dufrenite, variscite, scorodite, quartz and calcite.

The data obtained by the oscillation method offer confirmation of the interpretation of the Laue photographs, demonstrating that one inclined axis is present, but only one. This is shown in Fig. 3, where certain intensity differences are noticeable. The photographs produced by oscillation about the b axis were symmetrical, except for spots obviously caused by smaller adhering crystals and inclusions.

TABLE 1. POWDER DIAGRAM OF PHOSPHOSIDERITE
(Unfiltered Fe radiation— $r=57.3$ mm.)

No.	Indices	$d_{\text{calc.}}$	$d_{\text{exp.}}$	$I_{\text{exp.}}$	No.	Indices	$d_{\text{calc.}}$	$d_{\text{exp.}}$	$I_{\text{exp.}}$
1	<i>Imp.</i>	—	7.12	$\frac{1}{2}$	35	{ 310	1.739	1.739	$\frac{1}{2}$
2	<i>Imp.</i>	—	6.75	$\frac{1}{2}$		{ 301	1.735		
3	011	6.49	6.48	1		{ 134	1.713		
4	<i>Str.</i> ₆	—	5.37	$\frac{1}{2}$	36	{ 311	1.708	1.716	3
5	β 110	5.13	5.13	1		{ 015	1.707		
6	020	4.90	4.88	1	37	{ 152	1.693	1.691	1
7	β 002	4.78	4.81	1		{ 152	1.689		
8	110	4.66	4.67	5	38	{ 025	1.673	1.670	3
9	101	4.54	4.55	$\frac{1}{2}$		{ 320	1.662		
10	{ 101	4.50	4.48	$\frac{1}{2}$	39	{ 321	1.635	1.633	2
	{ β 111	4.50				{ 060	1.632		
11	002	4.335	4.327	6	40	{ 115	1.630	1.603	$\frac{1}{2}$
12	111	4.090	4.089	2		{ 321	1.629		
13	{ 012	3.964	3.965	2	41	{ 312	1.608	1.580	2
	{ β 120	3.961				{ 061	1.604		
14	<i>Str.</i> ₂	—	3.696	$\frac{1}{2}$	42	{ 224	1.580	1.555	2
15	120	3.596	3.592	5		{ 250	1.575		
16	121	3.313	3.320	2	43	{ 160	1.559	1.528	3
17	{ 031	3.054	3.052	3		{ 330	1.554		
	{ β 013	3.053			{ 153	1.553			
18	β 210	2.818	2.831	1	44	{ 331	1.532	1.506	1
19	{ 130	2.779	2.771	>10		{ 243	1.532		
	{ 013	2.772			{ 035	1.531			
20	200	2.650	2.651	1	45	{ 331	1.527	1.492	1
21	210	2.558	2.557	5		{ 303	1.515		
22	<i>Imp.</i>	—	2.502	$\frac{1}{2}$	46	{ 303	1.500	1.453	1
23	{ 211	2.448	2.452	1		{ 234	1.498		
	{ 113	2.447			{ 313	1.497			
24	{ 132	2.334	2.338	2	47	{ 234	1.486	1.439	1
	{ 220	2.331				{ 205	1.458		
25	{ 123	2.262	2.256	2	48	{ 323	1.447	1.406	1
	{ 202	2.251				{ 006	1.445		
26	140	2.222	2.221	2	49	{ 205	1.444	1.392	3
27	{ 014	2.116	2.123	3		{ 215	1.442		
	{ 222	2.062			{ 323	1.434			
28	{ 230	2.057	2.064	1	50	{ 341	1.411	1.376	3
	{ 104	2.013				{ 154	1.404		
29	{ 133	2.009	2.011	5	{ 154	1.399	1.376	3	
	{ 024	1.982			{ 106	1.398			
30	{ 142	1.981	1.986	1	51	{ 225	1.397	1.376	3
	{ 051	1.910				{ 106	1.390		
31	{ 213	1.906	1.912	1	52	{ 260	1.389	1.376	3
	{ 150	1.837				{ 253	1.387		
32	{ 223	1.806	1.813	2	53	{ 253	1.379	1.376	3
	{ 034	1.805				{ 304	1.376		
33	{ 143	1.767	1.766	3	54	{ 333	1.375	1.376	3
	{ 241	1.763							

The absence of (010) was observed for odd orders through the 7th, beyond which the presence or absence could not be ascertained with certainty. However, it is noteworthy that (001) first appears odd with appreciable intensity in the 9th order although it is possibly also present in the 7th, and (100) does not occur in an odd order until the 5th. If the absence of (010) in odd orders is characteristic the possible space groups are C_2^2 and C_{2h}^2 , because the cell is simple monoclinic. If (010) is not characteristically absent in odd orders the possibilities must be extended to include C_s^1 , C_2^1 and C_{2h}^1 .

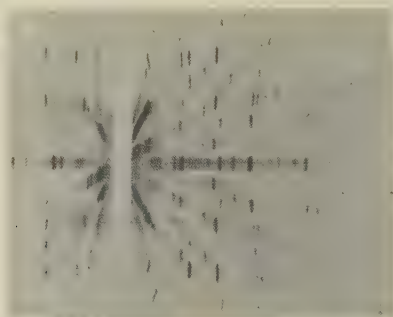


FIG. 3. Oscillation photograph produced by rotation of 20° about the c axis from 100, showing asymmetric intensities of the reflections.

In an attempt to limit further the space group, a number of crystals were etched on (010). The results obtained by etching with HCl were not gratifying. In general these efforts suffered from various sorts of interferences and figures which seemed to indicate planar symmetry were obtained as well as those which indicated axial symmetry. In addition some of the figures appeared to be asymmetric.

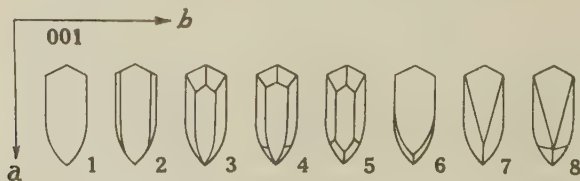


FIG. 4. Etch figures obtained by action of hot HCl on 001 (Redrawn from De Angelis).

The etch figures which De Angelis obtained seem to be more diagnostic than any that were obtained in this study and these are given in Fig. 4. It is highly probable that the symmetry of phosphosiderite is holohedral

because the figures obtained on 001 and $00\bar{1}$ were centrosymmetric, and there can be little doubt concerning the planar symmetry of these figures.

If it can now be concluded that the symmetry is holohedral and that (010) reflections are absent in the higher odd orders (above the 8th order) as well as in the lower orders the space group is C_{2h}^2 . If the symmetry is not holohedral C_2^2 is also a possibility, and if the odd high order reflections of (010) are not absent, C_{2h}^1 must be considered. Nevertheless, the space group is limited to five possibilities through consideration of the reflections which are definitely present.

The most probable formula that has been suggested for phosphosiderite is $\text{FePO}_4 \cdot 2\text{H}_2\text{O}$. Using this as a basis for calculating the density gives

$$\rho = \frac{4 \times 1.649 \times 186.9}{5.30 \times 8.67 \times 9.79 \times \sin 89^\circ 24'} = 2.74$$

The only recorded value for the specific gravity is 2.76, which agrees sufficiently well with the *x*-ray measurements and enhances the probability of this formula.

W. H. Zachariasen (8) was able to obtain an unique solution for a structure having the space group C_{2h}^2 and requiring the determination of sixteen parameters, but the analysis was irrelevant and involved a consideration of the intensities of some 700 reflections. The photographs obtained in the present study will not permit the assignment of indices and intensities to a large number of reflections. Under these circumstances it does not seem feasible to determine the structure by any straightforward method.

SUMMARY

Phosphosiderite is monoclinic and probably holohedral, in which case the space group is C_{2h}^2 or C_{2h}^1 . The apparent absence of (010) in odd orders recommends C_{2h}^2 . The value $\beta = 89^\circ 24'$ obtained by goniometric measurement (1) agrees excellently with the *x*-ray data. Other lattice constants are: $a_0 = 5.30$, $b_0 = 9.79$ and $c_0 = 8.67$ (all ± 0.01 Å). The unit cell contains four molecules of $\text{FePO}_4 \cdot 2\text{H}_2\text{O}$, yielding a theoretical density 2.74.

A powder diffraction pattern has been interpreted through the aid of Laue and oscillation photographs, and these measurements should prove useful in ascertaining which minerals are isostructural with phosphosiderite. The unfavorable geometry impedes a determination of the structure from the data obtainable.

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MICROCLINE IN THE NATIVE COPPER DEPOSITS OF MICHIGAN

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INTRODUCTION

Despite decades of intensive mining activity in the copper country of Michigan, and detailed study of the native copper deposits during this period, the fact that a certain feldspar abundant in several of the ore bodies is a very unusual variety of microcline, has not been appreciated. In this paper its distribution, nature and paragenesis are discussed. This information the writer believes will contribute to our knowledge of the native copper deposits and of that important mineral group—the potash feldspars.

SOME PREVIOUS OBSERVATIONS ON THE FELDSPARS OF THE NATIVE COPPER DEPOSITS

In the middle of the last century Foster and Whitney¹ were surprised at the “anomalous occurrence” in the newly discovered native copper deposits of Keweenaw peninsula of a “peculiar reddish variety of orthoclase” which they considered an “undoubted aqueous product” forming after both copper and calcite. Whitney² supplemented this with a complete analysis³ and further remarks on the feldspar’s widespread development in small amounts “in all the mines from Keweenaw Point to Ontonagon” in association with analcite, as the “last formed of all the vein minerals.” These early mines exploited vertical vein-like ore bodies, known locally as “fissures,” which transect nearly perpendicularly the lavas underlying the region. With the opening of mines in the “lodes,” as the mineralized vesicular and flow-brecciated tops of the basaltic flows and interbedded felsite conglomerates are called, a red feldspar quite different in appearance from the “orthoclase” found by the aforementioned workers became conspicuous as a gangue mineral. Pumpelly⁴ noticed that some amygdules contained a “red feldspar” intimately associated with prehnite. Julien⁵ described brick-red “orthoclase” lining calcite-filled vugs in amygdaloidal basalt at the outcrop of the ore body

¹ Foster, J. W. and Whitney, J. D., Report on the geology of the Lake Superior Land District: *Exec. Doc.*, No. 4, Pt. 2, 102 (1851).

² Whitney, J. D., Notice of new localities, and interesting varieties of minerals in the Lake Superior region . . . : *Am. Jour. Sci.*, **28**, 16–19 (1859).

³ SiO₂ 65.45, Al₂O₃ 18.26, Fe₂O₃ 0.57, K₂O 15.21, Na₂O 0.65 %.

⁴ Pumpelly, R., The paragenesis and derivation of copper and its associates in Lake Superior: *Am. Jour. Sci.*, 252 (1871).

⁵ Julien, A. A., Note on a feldspar from the Calumet Copper mine: *Annals N. Y. Acad. Sci.*, **12**, 650–654 (1900).

on which the Calumet Mine was sunk, pointing out its adularia-like habit and ferruginous nature.

In more recent times Palache and Wandke investigated the regional mineralogy. Their observations were incorporated in an exhaustive geologic study of the district.⁶ They explained that "orthoclase" occurs here in two forms—as a "red feldspar" which was one of the earliest minerals of the ore forming period and is abundant in several of the lodes, and as "adularia" which was a relatively late mineral and occurs in small amounts in lodes and fissures. Broderick,⁷ in an important contribution to the literature on the native copper deposits, considered "adularia" a characteristic mineral of the shallow and intermediate depth zones of these ore deposits. However, the term "adularia" was not used here for the late feldspar as in *Professional Paper 144*, but referred to the early red feldspar,⁸ which as will be demonstrated is microcline.

Although native copper deposits associated with basic extrusives are of world-wide distribution, the writer is aware of only two such localities outside of Michigan from which secondary feldspar is reported. A "pink feldspar" occurs in amygdules in the epidotized, schistose pre-Cambrian basaltic lavas in which the native copper deposits of the South Atlantic states are found,⁹ and "red feldspar" filling vesicles is associated with native copper in the traps of the Coppermine River region, Northwest Territory, Canada.¹⁰

THE DISTRIBUTION OF MICROCLINE IN THE NATIVE COPPER DEPOSITS

In the Lake Superior region this variety of microcline is confined to metallized portions of the Keweenaw series. In the southern part of the "Copper Range" microcline is a prominent gangue mineral in the lodes opened in the workings of the Mass, the Adventure, and the Michigan mines. About 60 miles northeast, the Arcadian and Superior lodes carry this mineral in abundance, and in the Pewabic lode and parts of the Isle Royale lode, but sparsely. Farther north microcline is characteristic of the mineralization in the Osceola lode and most of the Kearsarge lode, and is present but not abundant in the Calumet and Hecla conglomerate.

⁶ Butler, B. S., Burbank, W. S., et al, The copper deposits of Michigan: *U.S.G.S. Prof. Paper 144*, 59-60 (1929).

⁷ Broderick, T. M., Zoning in Michigan copper deposits and its significance, Pt. 2: *Econ. Geol.*, **24**, 311-326 (1929).

⁸ Personal communication, T. M. Broderick, January 1938.

⁹ Watson, T. L., The native copper deposits of the South Atlantic States compared with those of Michigan: *Econ. Geol.* **18**, 743 (1923).

¹⁰ Gilbert, G., Copper in the Coppermine River, N.W.T.: *Econ. Geol.*, **18**, 743 (1923)

The Baltic and the Isle Royale lodes which lie north and south, respectively, of the smaller and much less productive Superior lode are characterized by the abundant development of sericite and the complete, or almost complete, absence of microcline. In the Superior and Arcadian lodes both microcline and sericite are important gangue minerals.

Broderick¹¹ demonstrated that in the Kearsarge lode, which has been mined for miles along the strike and thousands of feet down the dip and hence was very suitable for detection of zoning, with increasing depth the mineralization changed progressively from a microcline ("adularia")—bearing type to a sericitic type. The relation observed by the writer between the microcline and sericite which occur in close association in the Arcadian and Superior lodes is significant because of the light it sheds on the process of ore deposition. In these ores sericite follows microcline in the paragenetic sequence. It is generally assumed that in hypogene ore deposits minerals are successively formed at gradually decreasing temperatures. This generalization is not wholly satisfying in view of the evidence for superposition of the mineralization characteristic of the deeper (and presumably hotter) zone on that of the shallower as indicated by the order of formation of sericite and microcline. The writer believes that during the period of mineralization leading up to the formation of sericite the succeeding minerals were deposited at increasing temperatures. Later minerals were probably precipitated from cooling solutions. There are also indications that there was a waxing and waning in the intensity of mineralization corresponding to the thermal variation.

THE NATURE AND PARAGENESIS OF THE MICROCLINE OF THE NATIVE COPPER DEPOSITS

The initial process in the formation of the ore deposits was the introduction of potash-rich solutions (source?) into the northwestward-tilted Keweenaw series. The hot alkaline solutions ascended for thousands of feet along the vesicular and flow-brecciated ferruginous lava tops leaching lime, soda and ferric oxide from the wall-rock and depositing microcline in vesicles, as irregular replacements, or as crusts on the walls of vugs formed where solution exceeded deposition.

The calcic plagioclase was especially amenable to attack. The opaque ferruginous groundmass was also replaced by microcline, but the abundant ferric oxide was not entirely removed, a share of it undergoing redistribution as finely divided hematite which imparts to the feldspar a bright brick-red color. The omnipresence of the inclusions of ferric oxide in the microcline wherever the latter is developed, whether replac-

¹¹ *Op. cit.*

ing minerals of little or no iron content, or filling voids, points to actual precipitation of hematite from solution contemporaneously with the feldspar.

Augite demonstrated a pronounced resistance to replacement by microcline. Its stability during this early stage in the mineralization is evinced by microscopically observable instances of the preservation of the ophitic texture whereby microcline pseudomorphous after labradorite came to lie enclosed in unaltered augite.

Wherever the microcline was deposited in open spaces it formed small (generally less than a few millimeters in size), usually singly-terminated, short-prismatic crystals with a small but significant variation in habit. The dominant forms of the earliest formed microcline are $m\{110\}$, $M\{\bar{1}\bar{1}0\}$, and $x\{\bar{1}01\}$. As crystallization proceeded $c\{001\}$ and $b\{010\}$ became conspicuous. In most crystals c and x are equally developed, which gives them an orthorhombic aspect. It is of utmost interest that the morphogenetic classification of the potash feldspars presented by Kalb¹² permits of direct comparison. The earliest crystals are similar to his rhombohedral-like low temperature hydrothermal form (such as occurs in epithermal gold deposits of the North American Cordilleran region), and the latest to his moderate temperature form (as exemplified by the adularia of the Alpine clefts). Thus it appears that even during the comparatively small part of the period of mineralization represented by the time in which microcline was deposited, the effect of rising temperature expresses itself by the variation in the morphology of the microcline. Whenever the side-pinacoid is present a clear inclusion-free zone paralleling the prism faces is conspicuous.

The complex twinning divulged when the apparently simple crystals are viewed between crossed nicols is the distinctive feature of this feldspar and one which distinguishes it from ordinary microcline of granites and associated pegmatites. Sections of crystals transecting the prism zone at steep angles to the c -axis demonstrate between crossed nicols a division into four sectors roughly delineated by the diagonals of the rhombus outlined by the traces of the prism faces. The diagonally opposite sectors extinguish simultaneously when the short diagonal or trace of the b cleavage is inclined at an angle of about 17° to the plane of vibration of the polarizer or analyzer. The central portion is an irregular interpenetration of the two phases. In many crystals the front- and side-pinacoids are composition planes near the vertices.

If the interpenetration type of twinning were subdued and the com-

¹² Kalb, G., Die Kristalltracht des Kalifeldspates in minerogenetischer Betrachtung: *Centr. Min.*, 454 (1924).

position planes better developed it would be similar to the intercrossing twins ("Durchkreuzungszwillinge") according to the albite law as first described by Rose¹³ for albite from Roc Tourné in Savoy and later by Lacroix¹⁴ for albite from the Pyrenees. In fact some of the microcline crystals approach this state very closely, but they are preponderantly a combination of intercrossing ("durchkreuzung") and interpenetration twins according to the albite law.

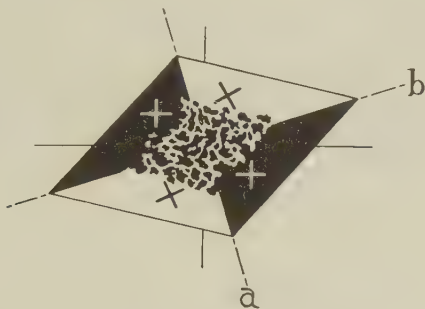


FIG. 1. Diagrammatic sketch showing the twinning of the microcline.

The irregular interpenetration twinning demonstrated by the microcline of the copper deposits has been described as characteristic of the perthitic microcline frommiarolitic cavities in the Riesengebirge and Striegau granites in Silesia,¹⁵ and the Waldstein granite of Fichtelgebirge.¹⁶ The writer has found microcline from a vug in a pegmatite at Pala, California, showing similar twinning, and is inclined to believe that further study may show pneumatogenic and hydrogenic microcline to be commonly of this nature.

A quantitative chemical analysis for potash and soda of a sample of microcline from the Michigan mine, Rockland, Ontonagon County, made by Dr. R. B. Ellestad of the Rock Analysis Laboratory of the University of Minnesota showed a K_2O content of 14.98% and a Na_2O content of 0.17%. Insofar as the writer is aware this extraordinary freedom from soda is not attained in any other potash feldspar, the analysis of which is presented in the literature, although data for orthoclase from

¹³ Rose, G., Ueber die Krystallform des Albits von dem Roc Tourné und Bonhomme in Savoyen—*Pogg. Ann. Phys. Chem.*, **125**, 459 (1865).

¹⁴ Lacroix, A., Notes sur quelques minéraux français. 1. Albite de Pouzac (Haut Pyrenées): *Bull. Soc. franc. Min.*, **11**, 70-71 (1888).

¹⁵ Beutell, A., Beiträge zur Kenntniss der schlesischen Kalinatronfeldspate: *Zeits. Krist.*, **8**, 352 and Fig. 1, Plate 6 (1884).

¹⁶ Durrfeld, V., Die Drusenmineralen des Waldsteingranits im Fichtelgebirge: *Zeits. Krist.*, **46**, 569 and Fig. 1b, Plate 12 (1909).

epithermal precious metal deposits and from the iron ores of Michigan¹⁷ approach it closely.

Pumpellyite, epidote, and perhaps chlorite overlapped slightly in period of deposition with the microcline. However, the bulk of these minerals was formed later than the feldspar, as is also the case for prehnite, sericite, quartz, calcite, datolite and copper.

In the North Kearsarge amygdaloid minute sharply twinned albite crystals occur implanted on microcline lining vesicles, or in some cases where the potash feldspar is totally deficient, albite, either clear or rendered cloudy by iron oxide inclusions, appears to be the earliest mineral of the ore-forming period and is engulfed and embayed by the later minerals, such as quartz, calcite and chlorite.

At the stage where KAlSi_3O_8 and Fe_2O_3 ceased to be precipitated the initial loci of mineralization (fractures in the lava top) had been broadened into small irregular vugs crusted with ferruginous microcline, or into drusy veinlets of the same substance. Wall rock alteration was insignificant, being confined to narrow zones of incipient replacement of the labradorite and ferruginous groundmass adjacent to the veinlets and nests of microcline. In the succeeding stage of mineralization the microcline-lined openings provided conduits for the chemically active ore-bearing solutions. The destruction of hematite inclusions, contrasted with its earlier deposition, was a dominant process. The microcline crystals were corroded and the hematite inclusions leached therefrom. At the same time precipitation of saturated phases in the voids and widespread metasomatic alteration of the wall rock prevailed. Pumpellyitization, epidotization, chloritization and silicification often extended far beyond the narrow confines of the feldspathization, and in such cases only vestiges of this phase of the mineralization persisted. In other portions of the lodes this stage of the mineralization was milder, being confined to filling and replacement within the feldspathized areas.

At the close of the period of calcite and copper deposition almost all the primary and secondary cavities in the intensively mineralized areas were tightly filled, but locally vugs lined with calcite or quartz crystals, or more rarely, with sheafs of prehnite, were still open in the now highly altered wall rock. In certain areas where the main part of the ore-forming period manifested itself least emphatically, vesicles lined with microcline persisted in but slightly altered basalt. The comparative weakness of the post-copper mineralization is attested by the almost complete restriction of mineral formation to the deposition of saponite, analcite and orthoclase along the walls of the relatively few open spaces which remained.

¹⁷ Classon, E., Mineralogical Notes: *Am. Jour. Sci.*, **23**, 67 (1882).

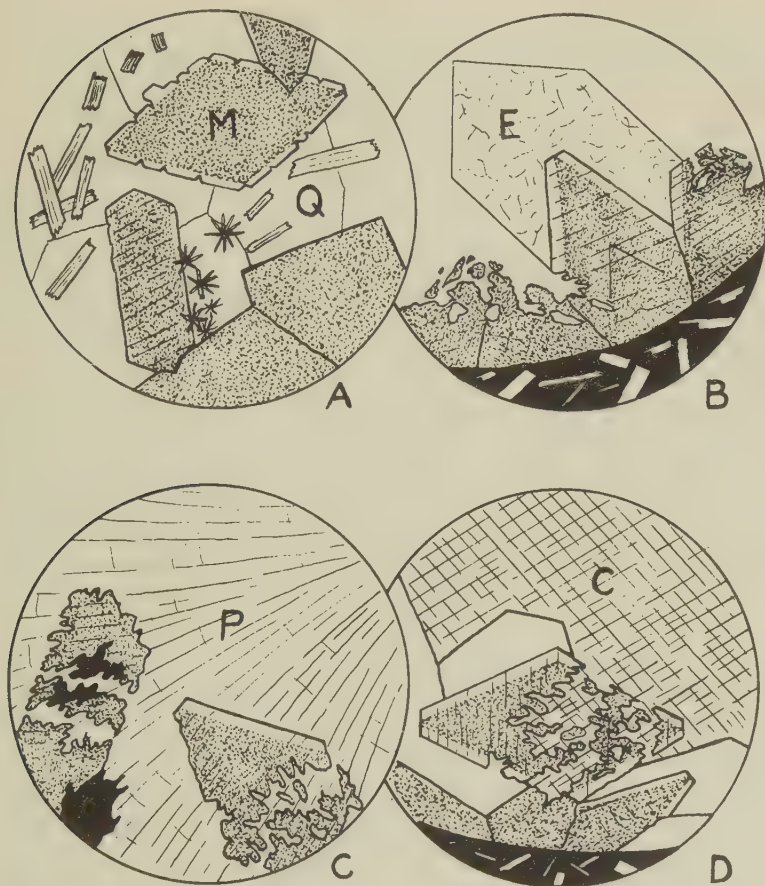


FIG. 2. Some features of the paragenesis seen in thin sections.

Magnification about 25X

A. Microcline (*M*) cloudy with hematite inclusions (stippled) engulfed by quartz (*Q*) and pumpellyite in groups of radiating acicular crystals and stouter prismatic individuals. Mass mine, Ontonagon County.

B. Portion of an amygdale showing the following order of mineral deposition: microcline and hematite, epidote (*E*), and quartz. Inclusion-free zone parallels the prism faces of the microcline. Mass mine, Ontonagon County.

C. Microcline embayed by prehnite (*P*), and both replaced by native copper (black). Sometimes prehnite is incipiently or entirely replaced by sericite with good preservation of the sheaf-like structure. Superior mine, Houghton County.

D. Portion of an amygdale showing the following order of mineral deposition: microcline and hematite, quartz, and calcite (*C*). Superior mine, Houghton County.

The orthoclase, in minute crystals of rhombohedral-like habit ($m\{110\}$ and $x\{\bar{1}01\}$), behaves very erratically in polarized light, showing wavy extinction and division into irregular sectors. Orthoclase from the gold telluride ore of Cripple Creek¹⁸ is reported to act similarly. The writer believes these phenomena are due to imperfect crystallization such as would result from compounding by subparallel growths. From veins at Rawhide, Nevada, Rogers¹⁹ described "valencianite" (vein orthoclase) displaying interesting optical anomalies which are suggestive of the twinning of the microcline of the copper deposits. He notes that some of the crystals in polarized light are divided into four sectors, which extinguish in diagonally opposite pairs, with extinction angles measured from the trace of the side-pinacoid of from 5 to 7°.

CONCLUSION

The "red feldspar" of the native copper deposits of Michigan is an unusual variety of microcline of low, or moderate temperature hydrothermal origin characterized by an adularia-like habit, intercrossed and interpenetrant twinning on the albite-law, and a very low soda content.

ACKNOWLEDGMENTS

The research which forms the basis of this paper was for the most part conducted by the writer as a student at the Michigan College of Mining and Technology, which institution provided funds for the preparation of thin sections and for a partial chemical analysis, and completed at Stanford University. Dr. T. M. Broderick generously placed the large collection of thin sections and accompanying notes of the Geology Department of the Calumet and Hecla Consolidated Copper Company at the writer's disposal. The writer is further indebted to Mr. R. N. Denning for valuable assistance in the collection of materials, and to Dr. A. F. Rogers, Dr. R. M. Dickey, and Dr. D. M. Lemmon for helpful suggestions during the course of the investigation and preparation of the manuscript.

¹⁸ Lindgren, W., and Ransome, F. L., Geology and gold deposits of the Cripple Creek district: *U.S.G.S. Prof. Paper* 54, 127 and 187 (1906).

¹⁹ Rogers, A. F., Orthoclase-bearing veins from Rawhide, Nevada, and Wechawken, New Jersey: *Econ. Geol.*, 6, 794 (1911).

UNIT CELL AND SPACE GROUP OF MONAZITE, (La,Ce,Y)PO₄

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Gordon has recently described monazite from Llallagua, Bolivia¹ and he supplied the writer with crystals for an equi-inclination Weissenberg study.² During the preparation of this manuscript Gliszczynski³ published data on monazite which he obtained from oscillation and rotation photographs; he apparently used the data listed in Klockmann's *Lehrbuch*² to aid him in indexing the photographs. The writer's results agree with those of Gliszczynski and are summarized below.

A description of the crystals used in this study is given by Gordon.¹ They are "translucent, and pink-flesh (Ridgway) in color. . . . The indices of refraction as determined by the oil immersion method are: $\alpha = 1.785$, $\beta = 1.787$, $\gamma = 1.840$; all ± 0.005 ." A chemical analysis is given below (Table 2).

Using unfiltered CuK radiation, rotation, 0-, 1st-, and 2nd-layer photographs around the *b*-axis, and rotation and 0-layer photographs around the *c*-axis were taken. A study of these photographs showed that monazite has a primitive monoclinic cell and that the space group is $P2_1/n$ (C_{2h}^5). The crystal class is therefore uniquely determined as C_{2h} .

Refined measurements of the lattice constants were not attempted because the crystal used was rather large and gave intense Weissenberg background patterns.⁴ In addition, there were only a few pinacoid reflections in the high- θ regions. The lattice constants listed below (Table 1) are based on measurements of four high- θ axial zone reflections; *a* and *c* were computed from the calculated β angle.

¹ Gordon, Samuel G., Thorium-free monazite from Llallagua, Bolivia: *Notulae Naturae, Acad. Nat. Sci. Phila.*, No. 2, May 17, 1939.

² No *x*-ray data on monazite appears in *Strukturbericht*. The writer was not aware of the data listed in Klockmann's *Lehrbuch der Mineralogie*, 11th ed. by P. Ramdohr, Stuttgart, 1936, p. 460. No reference to the original paper is given.

³ Gliszczynski, S. von, Beitrag zur "Isomorphie" von Monazit und Krokoit: *Zeits. Krist. (A)*, **101**, 1-16 (1939).

⁴ Buerger, M. J., X-ray surface reflection fields and their application to absorption corrections and to background patterns: *Zeits. Krist. (A)*, **99**, 189-204 (1938).

TABLE 1

Gliszczynski		Parrish
6.782	<i>a</i>	6.76
6.993	<i>b</i>	7.00
6.445	<i>c</i>	6.42
76°22'	β	76°50'
0.9698:1:0.9231	<i>a:b:c</i>	0.9660:1:0.9167
	Sp. gr., meas.	5.173
5.217	Sp. gr., calc.	5.06
	Cell volume	296

TABLE 2

	1.	2.	3.	4.
P ₂ O ₅	29.29	29.59	0.208	9.04
SiO ₂	0.27			
Ce ₂ O ₃ , etc.,	31.41	31.74	0.097	4.22
La ₂ O ₃ , etc.,	33.19	33.54	0.094	4.09
Y ₂ O ₃ , etc.,	5.08	5.13	0.023	1.00
CaO	0.34			
MgO	0.22			
	99.80	100.00		

1. Chemical analysis by Samuel G. Gordon (1). A spectrographic analysis showed no thorium.

2. Analysis recalculated to 100% after deducting SiO₂, CaO, MgO.

3. Molecular ratio.

4. Molecular ratio on basis of Y₂O₃=1.00.

This data yields the formula $4\text{La}_2\text{O}_3 \cdot 4\text{Ce}_2\text{O}_3 \cdot \text{Y}_2\text{O}_3 \cdot 9\text{P}_2\text{O}_5$ which is equivalent to $\text{La}_4\text{Ce}_4\text{YPO}_4$ or $(\text{La}_{4/9}, \text{Ce}_{4/9}, \text{Y}_{1/9})\text{PO}_4$ which agrees with the general monazite formula $(\text{La}, \text{Ce}, \text{Y})\text{PO}_4$. There are four molecules of $(\text{La}_{4/9}, \text{Ce}_{4/9}, \text{Y}_{1/9})\text{PO}_4$ per unit cell.

The writer wishes to thank Prof. M. J. Buerger for his aid and valuable suggestions in this problem, and Mr. Samuel G. Gordon of the Academy of Natural Sciences, Philadelphia, for his cooperation in supplying the crystals.

SOME MODES OF QUARTZ-BEARING PLUTONITES FROM DERBY, VERMONT

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INTRODUCTION

Derby township, the southern part of which was formerly known as Salem, is situated in the north central part of the State of Vermont. It is bounded on the west by lake Memphremagog and on the north by the province of Quebec. Rocks of the granitic type have long been known to occur in this district.¹ With very few exceptions, and upon megascopic examination only, they were simply designated as granites. Dale,² from a qualitative study, called the rock from the Newport quarry a quartz-monzonite.

No modal analyses of these rocks were found in the literature and since many specimens from the area were available among Dr. Charles H. Richardson's large collection of Vermont rocks, the author thought it might be of value to study them in more detail so that they could be classified on a quantitative basis and a more detailed petrographic description given of them. The modes of the rocks only are given in this paper; it is hoped that the detailed petrographic study will be available for publication in the near future.

The modes of the various rocks were determined by the Rosiwal method and were assigned to the various families, orders and classes in accordance with Johannsen's classification.³ A few of the assigned names may be changed with further study, especially where the plagioclase-potash feldspar ratio is close to the dividing line between two families, as in No. 2, where an increase of less than one per cent in the microcline content would change the rock from a tonalite to a granodiorite. Likewise, where the biopyribole ratio is near the boundary between two classes, as in No's. 5 and 10, a small increase in the biotite content would indicate a granodiorite rather than a leucogranodiorite. For convenience and comparison the modes of the various rocks are grouped together in table 1.

¹ Hitchcock, C. H., The Geology of Northern New England, (1870-1882), 2-6. Richardson, C. H., Areal and Economic Geology of North Western Vermont, *Rept. Vt. State Geologist* (1905-1906), 106-109; The Geology of Newport, Troy and Coventry, *Rept. Vt. State Geologist* (1907-1908), 280. Dale, T. Nelson, Commercial Granites of New England, *U. S. Geol. Survey Bull.* **738**, (1923), 119-120; The Granites of Vermont, *Rept. Vt. State Geologist* (1909-1910), 113-114.

² Dale, T. Nelson, *ibid.*, 119-120.

³ Johannsen, Albert, A Descriptive Petrography of the Igneous Rocks, **I**, 140 (1931).

TABLE 1. MODES OF QUARTZ-BEARING PLUTONITES

	1	2	3	4	5	6	7	8	9	10
Quartz	23.64	18.26	20.18	21.40	22.68	19.00	22.18	13.01	11.78	41.48
Microcline	3.42	4.49	13.82	17.12	12.89	8.51	13.21	17.50	25.50	13.50
Oligoclase	68.46	69.20	52.10	45.72	55.20	65.98	57.80	57.42	49.21	39.67
Muscovite	x	0.75	1.95	4.94	4.73	0.43	x	1.01	0.51	x
Apatite	x	x	x	x	x	x	x	x	x	x
Biotite	4.06	6.88	11.70	9.88	4.50	6.06	6.13	11.26	13.00	4.96
Zircon	x	x	x	x	x	x	x	x	x	x
Sphene		x	x		x	x		x		
Orthite				x						x
Epidote	x	0.50		0.36	x		x	x		x
Chlorite				x	x			x	x	
Calcite	x					x		x		
Kaolinite	x	x	x	x	x	x	x	x	x	x
Secondary										
mica	x	x	x	x	x	x	x	x	x	x
Garnet										x
Rutile	?	?	?	?	?	?	?	?	?	?

x The mineral is present.

? May be present but not certain of the identification.

1. Leucotonalite, Johannsen, (128P). La Casse quarry, Derby, Vermont; C. H. Richardson's collection, "Granite."
2. Biotite-tonalite, Johannsen, (228P). South of La Casse quarry, Derby, Vermont; C. H. Richardson's collection, "Granite."
3. Biotite-granodiorite, Johannsen, (227P), Newport quarry, Derby, Vermont; T. Nelson Dale, U. S. Geological Survey, Bull. 738 (1923), pp. 119-120. "Quartz monzonite."
4. Biotite-granodiorite, Johannsen, (227P). Parameter quarry, Derby, Vermont; *Idem*: p. 120. "Granite" or "quartz monzonite."
5. Leucogranodiorite, Johannsen, (127P), Beebe Plains, Derby, Vermont; C. H. Richardson, Building Stones and Clays, 1917, p. 83, "Granite."
6. Biotite-granodiorite, Johannsen, (227P), Derby Line, Derby, Vermont; C. H. Richardson's collection, "Granite."
7. Biotite-granodiorite, Johannsen, (227P), Salem Mountain, Derby, Vermont; C. H. Richardson's collection, "Granite."
8. Biotite-granodiorite, Johannsen, (227P), South side of Salem Mountain, Derby, Vermont; C. H. Richardson's collection, "Granite."
9. Biotite-granodiorite, Johannsen, (227P), Clyde River, Salem, Vermont; C. H. Richardson's collection, "Granite."
10. Leucogranodiorite, Johannsen, (127P), "West side exposures in center of Salem township"; Vermont; C. H. Richardson's collection, "Granite."

GENERAL DISCUSSION

The exact geological age of these rocks is in doubt. According to Jacobs⁴ the youngest rocks that they intrude in Vermont are Ordovician. Just north in Quebec, however, similar rocks are found intruding Devonian strata, which would make them post-Devonian in age. Jacobs⁵ further states that some geologists go so far as to put the time of intrusion as post-Carboniferous. None of the rocks examined in this study showed any evidence of having suffered dynamic metamorphism such as might have been brought about by the Taconic disturbance at the end of the Ordovician. Thus this study only substantiates the conclusion that they are at least post-Ordovician in age.

The most outstanding fact that is brought out by a study of table 1 is the complete absence of orthoclase feldspar. This paucity, however, is not characteristic of this area of Vermont alone, for with the possible exception of the West Dummerston leucogranodiorite, the author⁶ found practically no orthoclase in any of the quartz-bearing plutonites that he studied covering many widely scattered areas throughout the State. Even the small amount of orthoclase listed for this West Dummerston area is undoubtedly too high for a very detailed study (82 sections) by Church⁷ lists orthoclase as "little." Probably the few grains listed by both authors are really microcline in which the characteristic twinning does not show, for Johannsen⁸ states "in some cases the twinning in microcline becomes of such microscopic dimensions that it is recognizable only in the thinnest sections—in still other cases there appears to be no twinning and microcline can be distinguished from orthoclase only by its extinction angle of 15° – 16° on (001)." In thin sections where there are only one or two doubtful grains it would indeed be exceptional if any of them were so orientated as to give a correct determination on extinction alone. Examination of many thin sections of these post-Devonian plutonic rocks from the Appalachian system even as far south as Georgia shows the same remarkable lack of orthoclase.

Table 1 also shows oligoclase to be the most abundant constituent in any of the rocks. A great many refractive index determinations showed

⁴ Jacobs, Elbridge C., *An Account of Vermont Geology, Rept. VI. State Geologist* (1935-1936) 137.

⁵ *Ibid.*, 137.

⁶ Maynard, J. E., The petrographic re-examination of quartz-bearing plutonites from Vermont: *Jour. Geol.*, **XLII**, 146-162 (1934).

⁷ Church, Mary S., A quantitative petrographic study of the Black Mountain leucogranodiorite at West Dummerston, Vermont: *Jour. Geol.*, **XLV**, 763-774 (1937).

⁸ Johannsen, Albert, *A Descriptive Petrography of the Igneous Rocks: II*, 145-146 (1932).

its composition to vary between Ab 76 and Ab 84. Mulholland⁹ made a chemical analysis of the plagioclase which he separated from the most typical granodiorite of the Derby area. His results yielded oligoclase of the composition Ab 76 An 24.

Another notable fact is the relatively low percentage of potash feldspar as represented by microcline, the average for the ten rocks being 13.0 per cent, in contrast to, the relatively high percentage of oligoclase, the average in this case being 56.0 per cent. These figures and the general petrographic study suggest that the township of Derby at least, and perhaps a considerable part of the Appalachian system, was underlain by granodiorite magma or magmas of which the potash feldspar-plagioclase ratio was such as to yield granodiorite batholiths the composition of which were much nearer tonalites than quartz-monzonites. This expresses itself locally in the actual occurrence of tonalites as differentiates from the main granodiorite magma or magmas. The great mass of this batholith or batholiths has yet to be exposed by erosion, the present exposures being stocks or bosses that represent cupolas of the main masses.

⁹ Mulholland, M. M., Zoning as an explanation of optical anomalies of a plagioclase feldspar in quartz-bearing plutonites from Vermont: *Am. Mineral.*, **23**, 534-536 (1938).

BOOK REVIEW

AN INTRODUCTION TO CRYSTAL CHEMISTRY. R. C. EVANS. vii+388 pp., 59 Tables, 113 Figs. Cambridge, at the University Press. (New York: The Macmillan Company.) 1939. Price \$4.50.

This volume is designed primarily as a text for university students, but will be of interest to many others. No specialized crystallographic training is necessary for the reader. Very little attention is given to detailed structural or crystallographic features, nor is there any attempt to make a complete survey of all structures which have been analyzed. The main emphasis is on the correlation of typical structures with their chemical and physical properties. The various chapters consider interatomic binding forces, quantitative lattice theory, metallic elements, alloy systems, homopolar, ionic and molecular compounds. The material is clearly presented and well illustrated.

L. S. RAMSDELL

NEW MINERAL NAMES

Trieuite

L. DE LEENHEER: Trieuit, een nieuw kobaltmineraal. *Natuurw. Tijdschr.*, Nr. 4-5 91 (1935); Sur la Trieuite et les Minéraux associés. *Comité Spécial du Katanga; Ann. Serv. Mines*, 8, 3-13, 1937 (1938), 3 figs., 1 pl.

NAME: In honor of Robert du Trieu de Terdonck, Chief Geologist of the Union Minière du Haut-Katanga.

CHEMICAL PROPERTIES: A hydrated cobalt copper oxide: $2\text{Co}_2\text{O}_3 \cdot \text{CuO} \cdot 6\text{H}_2\text{O}$. Analysis: Co_2O_3 53.38, CuO 22.63, SiO_2 1.50, H_2O 20.16, CO_2 1.36. Sum 99.23.

PHYSICAL AND OPTICAL PROPERTIES: Black, luster vitreous and brilliant, streak black-brown, fracture conchoidal. $H=3.5$. $G=3.128$. In thin section brown, isotropic. $n=1.85$.

OCCURRENCE: As a secondary mineral associated with malachite and quartz at the Etoile du Congo mine, near Elisabethville, Belgian Congo. Formerly called heterogenite.

W. F. FOSHAG

Cupro-asbolane

L. DE LEENHEER: Sur quelques minerais de manganèse du Katanga. *Comité Spécial du Katanga, Ann. Serv. Mines*, 8, 32-64, 1937 (1938), 11 figs., 1. pl.

NAME: From its relationship to asbolane—a copper bearing asbolane.

CHEMICAL PROPERTIES: A hydrated manganese cobalt copper oxide, of variable composition. Analysis (from Katanga): Residue 6.88, CuO 20.79, Fe_2O_3 ($-\text{Al}_2\text{O}_3$) 5.04, MgO 0.90, Co_2O_3 13.40, Mn_2O_3 38.67, H_2O 11.51. (Analyses of cupro-asbolane from other Congo localities also given.) Easily soluble in hydrochloric acid with liberation of chlorine.

PHYSICAL PROPERTIES: Color black, streak black, soft. G . about 2.37.

OCCURRENCE: Found at Katanga, Ruashi, and other localities in belts of the Belgian Congo, associated with quartz, malachite, carnotite (at Ruwe) etc., the so-called "Black Ore" of this district.

DISCUSSION: The psilomelane group of minerals is considered essentially manganese dioxide with adsorbed oxides of other metals: with manganese oxide, psilomelane or wad; with cobalt, asbolane; with copper, lampadite; with lithium, lithiophorite.

Cupro-asbolane is an intermediate member with important amounts of both cobalt and copper.

W. F. F.

Sharpite

J. MÉLON: La sharpite, nouveau carbonate d'uranyle du Congo belge. *Bull. Inst. Roy. Colonial Belge*, 9, 333-336 (1938).

NAME: In honor of R. R. Sharp, who discovered the uranium deposits of Chinkolobwe in 1915.

CHEMICAL PROPERTIES: A hydrated uranium carbonate: $6\text{UO}_3 \cdot 5\text{CO}_2 \cdot 8\text{H}_2\text{O}$. Analysis: UO_3 81.04, CO_2 10.20, H_2O 6.81, CaO 2.70. Sum 100.85. Contained 1.6% insoluble matter. Dissolves easily with effervescence in dilute acids.

CRYSTALLOGRAPHICAL PROPERTIES: Orthorhombic (?).

PHYSICAL AND OPTICAL PROPERTIES: Color yellowish green. H = above $2\frac{1}{2}$. G > 3.33.

Extinction parallel, elongation positive. α = 1.633, γ near 1.72. Birefringence very high. Pleochroism feeble; X = brownish, Z = clear yellow—slightly greenish.

OCCURRENCE: Found as a radially fibrous crust with uranotile on a mass of curite and becquerelite.

W. F. F.

Teineite

TOYOHUMI YOSIMURA: Teineite, a new tellurate mineral from the Teine Mine, Hokkaido, Japan. *Jour. Faculty Science, Hokkaido Imp. Univ.*, Series IV, Nos. 3-4, Geol. & Mineral., 465-470 (1939), 7 figs.

NAME: From the Teine mine, where this mineral was first found.

CHEMICAL PROPERTIES: A hydrated tellurate-sulfate of copper. $10\text{CuTeO}_4 \cdot 3\text{CuSO}_4 \cdot 26\text{H}_2\text{O}$. Analysis: CuO 28.0, TeO_3 48.0, SO_3 6.6, H_2O 12.2, insol. 6.1. Sum 100.9. Soluble in HCl to a greenish yellow solution; in HNO_3 to a blue solution, first with separation of white telluric oxide, then to complete solution. In closed tube, gives off water. Fusible, 2, to a black bead.

CRYSTALLOGRAPHICAL PROPERTIES: Orthorhombic, prismatic. $a:b:c = 0.705:1:0.786$. $m \wedge m''' = 70^\circ 27' (\pm 5')$; $e \wedge e' = 76^\circ 20' (\pm 10')$. Forms: b (010), m (110), e (011), f (073).

PHYSICAL AND OPTICAL PROPERTIES: Color cerulean blue, often cobalt blue or bluish gray. Streak bluish white. Pleochroic X = greenish blue, Y = blue, Z = indigo blue. Absorption $Z > Y > X$. Biaxial, negative. $2V = 36^\circ$; plane of the optic axes parallel to (010); $X = a$, $Y = b$, $Z = c$. $\alpha = 1.767$, $\beta = 1.782$, $\gamma = 1.791$.

Cleavage (010) good; (001), (100) weak. $H = 2.5$, brittle, $G = 3.80$.

OCCURRENCE: Found in the oxidized portion of the Takinosawa vein that carries, besides pyrite, tetrahedrite and blende, tellurium and sylvanite, in a gangue of chalcadonic quartz and barite.

W. F. F.

Kasoite**Barium-albite**

TOYOHUMI YOSIMURA: Studies on the minerals from the manganese deposits of the Kaso Mine, Japan. *Jour. Faculty Science, Hokkaido Imperial University*, Series IV, Nos. 3-4, Geol. and Mineral., 313-453 (1939), 19 plates, 13 text figs.

CRYSTALLOGRAPHIC PROPERTIES: Crystal habit of the adularia type. Twinning absent.

CHEMICAL PROPERTIES: A barium feldspar: $\text{Ca}_{49.6}\text{Kp}_{32.5}(\text{Ab}, \text{Ne})_{17.9}$. Analysis: SiO_2 38.48, Al_2O_3 23.61, Fe_2O_3 0.60, MgO 0.97, MnO 2.67, CaO 0.85, BaO 25.50, Na_2O 1.85, K_2O 5.10, Ign. loss 0.98. Sum 100.61.

PHYSICAL AND OPTICAL PROPERTIES: Biaxial, negative, $2V=80^\circ$. Plane of the optic axes parallel to (010) (or nearly so). $c \wedge X$ on 010 $= 2^\circ-3^\circ$ in acute β ; $a \wedge Z$ on 011 $= 28^\circ-29^\circ$ in obtuse β . Extinction angle on (001) $= 10^\circ-13^\circ$. $\alpha=1.564$, $\beta=1.568$, $\gamma=1.572$. $G=3.003$. $H=5\frac{1}{2}$.

OCCURRENCE: Found as veinlets, with rhodonite or as impregnations in slate wall rock.

DISCUSSION: Kasoite is distinguished from other celsian feldspars by its high content of the kaliophilite and nepheline molecules. Other feldspar names: *Barium albite*, $Ab_{44} \cdot Or_{42} \cdot Cn_{14}$, found in manganhedenbergite veins.

W. F. F.

Iron knebelite
Manganknebelite
Iron tephroite
Picrotephroite

TOYOHUMI YOSIMURA: Studies on the minerals from the manganese deposits of the Kaso Mine, Japan. *Jour. Faculty Science, Hokkaido Imperial University*, Series IV, Nos. 3-4, Geol. and Mineral., 313-453 (1939), 19 plates, 13 text figs.

Names given for various members of the knebelite-tephroite series are:

Fayalite	100-95% Fe_2SiO_4
Manganfayalite	95-80
Iron knebelite	80-60
Knebelite	60-40
Manganknebelite	40-20
Iron tephroite	20- 5
Tephroite	5- 0

Of this series, those from iron knebelite to iron tephroite are found at the Kaso Mine. Also included is a magnesian member picroknebelite.

	Iron knebelite	Picro- knebelite	Mangan- knebelite	Iron tephroite
α	1.796	1.787	1.795	1.787
β	1.830	1.815	1.830	1.811
γ	1.845	1.830	1.840	1.819
$2V$	50°	56°	50°	60°
G	4.16	3.98	4.01	3.96
Fe_2SiO_4	57.0	26.3	24.0	16.3
Mn_2SiO_4	29.1	59.0	68.2	76.7
Ca_2SiO_4	4.5	4.2	0.6	—
Mg_2SiO_4	9.4	10.5	7.2	7.0

W. F. F.

Manganactinolite
Mangantremolite

TOYOHUMI YOSIMURA: Studies on the minerals from the manganese deposits of the Kaso Mine, Japan. *Jour. Faculty Science, Hokkaido Imperial University*, Series IV, Nos. 3-4, Geol. and Mineral., 313-453 (1939), 19 plates, 13 text figs.

Among the amphiboles of the Kaso Mine are two, given new terms:

	Manganactinolite	Mangantremolite
$\text{Fe}_2\text{Si}_2\text{O}_6$	26.8	18.3
$\text{Mn}_2\text{Si}_2\text{O}_6$	11.0	13.1
$\text{Ca}_2\text{Si}_2\text{O}_6$	26.7	25.3
$\text{Mg}_2\text{Si}_2\text{O}_6$	35.5	43.3
α	1.648	1.637
β	1.661	1.650
γ	1.668	1.660
2V	74°	84°
$c/\wedge z$	16°	15°

W. F. F.

Ca-Fe-spessartine**Fe-Ca-spessartine**

TOYOHUMI YOSIMURA: Studies on the minerals from the manganese deposits of the Kaso Mine, Japan. *Jour. Faculty Science, Hokkaido Imperial University*, Series IV, Nos. 3-4, Geol. and Mineral., 313-453 (1939), 19 plates, 13 text figs.

The garnet minerals of the Kaso Mine show some variation, two of which have been given distinct terms:

	Ca-Fe-spessartine	Fe-Ca-spessartine
$3\text{CaO} \cdot \text{Fe}_2\text{O}_3 \cdot 3\text{SiO}_2$	—	0.9
$3\text{CaO} \cdot \text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2$	23.7	29.6
$3\text{FeO} \cdot \text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2$	17.7	15.5
$3\text{MnO} \cdot \text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2$	57.4	51.7
$3\text{MgO} \cdot \text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2$	1.2	2.3
n	1.790	1.788
G.	3.83	3.97

W. F. F.

Ca-rhodochrosite**Fe-rhodochrosite****Fe-Mn-calcite**

TOYOHUMI YOSIMURA: Studies on the minerals from the manganese deposits of the Kaso Mine, Japan. *Jour. Faculty Science, Hokkaido Imperial University*, Series IV, Nos. 3-4, Geol. and Mineral., Vol. IV, 313-453 (1939), 19 plates, 13 text figs.

The carbonate minerals of the Kaso Mine include some intermediate members among which are the following:

	FeCO_3	MnCO_3	CaCO_3	MgCO_3	ω	G.
Fe-Mn-calcite	24.6	16.5	53.6	5.3	—	3.38
Ca-rhodochrosite	7.2	74.7	11.8	6.3	1.795	3.51
Ca-rhodochrosite	4.3	53.8	37.5	4.4	1.731	3.05
Fe-rhodochrosite	12.7	75.0	6.9	5.4	1.790	3.38

These carbonates occur with rhodonite and other manganese-iron silicates.

W. F. F.